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Counsellor of the Presidium of the Russian Academy of Sciences (RAS)
Address for correspondence:
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New Data on Silicified Microfossils from the Satka Formation of the Lower Riphean Stratotype, the Urals

V. N. Sergeev* and Lee Seong-Joo**

**Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 119017 Russia*

***Geological Faculty, Kyungpook National University, Taegu, 702-701 Republic of Korea*

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Abstract—Silicified microfossils found in the Satka Formation of the Burzyan Group (Lower Riphean stratotype of the southern Urals) are described. The research carried out augments the database on taxonomic composition of relevant microbiota. Besides cyanobacterial remains formerly known from that formation, abundant silicified microfossils representing eukaryotic microorganisms of phytoplankton are found. Representatives of these microorganisms occurring in a form of organic-walled remains in shaly siliciclastic rocks of the Satka and overlying Bakal formations have never been reported from cherty–carbonate successions of the Lower Riphean. In addition, the taxonomic composition of silicified cyanobacterial remains from the Satka Formation has been revised with due account for the current classification of fossil blue-green algae. The new finds and taxonomic correction of known forms elucidate further the microphytological characterization of the Lower Riphean and are important for biostratigraphic correlation of Precambrian deposits and for resolution of relevant problems. Described in the work are 15 species of 13 genera, the new form *Corymbococcus rexii* sp. nov. included.

Key words: microfossils, cyanobacteria, eukaryotic phytoplankton, Lower Riphean, standard Riphean succession, the Urals.

INTRODUCTION

Microphytological characterization of the Riphean stratotype in the Bashkirian meganticlinorium of southern Urals [1] is of a prime significance, because the microfossil succession established here represents a biostratigraphic standard for correlation of Precambrian microbiotas known in other regions. During the investigation history that lasted nearly fifty years, the microphytological characterization of the stratotype was substantiated based on organic-walled [2] and silicified microfossils [3]. The established successive assemblages of these organic remains suggest that there was a series of biotic events in the evolution of the Precambrian microworld. Nevertheless, the problems of stratigraphic correlation between assemblages from different regions and relevant disagreements in age determination, which are sometimes as great as two erathems (e.g., [4]), imply that it is necessary to get a deeper insight into the microphytological characterization of the Riphean stratotype. In addition, a considerable new paleontological material obtained in other regions during the last 15–20 years is quite stimulating from the viewpoint of microfossil biostratigraphy. On the other hand, the taxonomy of microfossils has been repeatedly revised during these years, and the formal correction of taxonomic compositions formerly reported for the Uralian microbiotas is desirable. All the aforesaid inspired the works aimed at the additional paleontological study of the Riphean stratotype in southern Urals and at the reexamination of its microbiotas described in earlier publications. During the fieldwork seasons of 1998–2000, cherty rocks were sampled in formerly known and new microfossil localities. This work is the first one in a series dedicated to microbiotas of three Riphean subdivisions. Presented in the work is new information on taxonomic composition and stratigraphic interpretation of silicified microbiota from the Satka Formation of the Lower Riphean studied in the Kusa area of the southern Urals.

GEOLOGICAL STRUCTURE OF LOWER RIPHEAN DEPOSITS IN THE KUSA AREA

The general stratigraphic succession of deposits in the Riphean stratotype area that was established in the 1930s is out of doubts to the present time [5]. Sedimentary rocks are divided here (from the base upward) into the Burzyan, Yurmata, and Karatau groups, each representing type sections of the Lower, Middle, and Upper Riphean, respectively, and corresponding to a large sedimentary cycle separated from the others by hiatuses and angular unconformities. In terms of structural geology, the Riphean deposits form the Bashkirian (Ural-tau) meganticlinorium divisible into fold structures of lower ranks. According to recent interpretations [6], the meganticlinorium is composed of a series of tectonic

slices, which form five large allochthonous units thrust eastward over each other.

The Burzyan Group (Lower Riphean). Sedimentary rocks of the group are exposed over the large northern and eastern areas of the Bashkirian meganticlinorium, and their sections are well correlated [7]. Within limbs of the Bakal–Taratash anticlinorium, northern sections of the Burzyan Group are divided into the Ai, Satka, and Bakal formations, the former of siliciclastic and the latter two of siliciclastic–carbonate deposits. In the arch area of that anticlinorium, there is exposed the Taratash Complex of pre-Riphean gneisses, crystalline schists, amphibolites, and quartzites. In the anticlinorium limbs, rocks of the complex are overlain by sediments of the Ai Formation, which enclose pebbles of underlying metamorphic rocks [8]. A borehole drilled in the western limb showed that at the depth of 600 m migmatites of the Taratash Complex are thrust over Paleozoic limestones [9]. The composition and structure of Lower Riphean successions are sometimes considerably different in the western Arsha and eastern Kisegan lithologic–structural zones of the anticlinorium.

The Satka Formation, rocks of which yield microfossils, is composed predominantly of dolostones, dolomitic limestones and marls with subordinate intercalations of limestones, marls, and carbonaceous sericite- or sericite–quartz phyllites. The formation is divided into the lower and upper Kusa, Polovinkino, lower and upper Satka subformations. Among these subformations, the former two are traceable in the Arsha zone, and the latter three in the Kisegan zone, where the rocks are altered to a greater extent [10]. The formation can be as thick as 2000–2400 m, although its thickness is considerably reduced in eastern areas.

The Satka Formation deposits bearing silicified microfossils have been repeatedly described in publications and are traditionally divided in the Kusa area into the lower and upper Kusa subformations [11].

The lower Kusa Subformation of pure, cherty, sandy, and marly dolostones is 700–900 m thick (this thickness decreases eastward). Among white compact fine-grained dolostones prevailing in the formation lower part, there are locally exposed horizons of highly silicified and stromatolitic dolostones with abundant beds and concretions of dark, early diagenetic. Pure dolostones of the subformation are associated with widespread sandy varieties, which grade into dolomitic sandstones.

The upper Kusa Subformation is represented by light to dark gray pure and sandy dolostones displaying locally the oolitic or pseudo-oolitic texture. These rocks are intercalated with beds of greenish sericite–chlorite–quartz, brown sericite–quartz, and greenish to dark gray sericite–carbonate slates. In the southern part of the Kusa area, the subformation is 500 to 800 m thick.

The study area, microfossils from which are described below, is of a complex tectonic structure, and relationships between the Satka Formation sections exposed in different localities should be described in detail. According to all publications mentioned above, a classical section of lower and upper Kusa subformations is exposed at the right bank of the Ai River near the railway bridge (Fig. 1). In fact, Riphean deposits of the Kusa area represent a nappe or a series of imbricated tectonic slices thrust over the Paleozoic deposits. In general, the structure resembles a monocline, although it is not very clear where we deal with the normal stratigraphic succession, and where with the structural succession of tectonic slices thrust over each other.

The basal tectonic slice of quartz sandstones is undoubtedly thrust over the Paleozoic rocks, because in some tectonic windows there are exposed presumably Devonian beds dipping at high angles. The sandstones age is problematic: in many works these rocks are attributed to the Upper Riphean Zilmerdak Formation [12], although they may also belong to different subformations of the Middle Riphean Zigal'ga Formation or to the Lower Riphean Ai Formation.

The slice of quartz sandstones is overridden by a thrust sheet composed of dolostones with cherts and intercalated shale and siltstone members. These rocks are exposed in several outcrops on the right side of the Ai River downstream of the Kusa Town (exposures U-1, Ulin Kamen Cliff, and U-2), where the best specimens of silicified microfossils have been collected (Fig. 1). It is remarkable as well that near the railway bridge the quartz sandstones grade into the fine-grained silty sandstones with glauconite and into plicated siltstones. In a close proximity, near the railway, there are exposed carbonate rocks with laminated stromatolites and cherts (exposure U-3, Fig. 1). Transitions between the rocks resemble in general a normal stratigraphic contact, for instance that between the Inzer and Min'yar formations of the Upper Riphean, although it is more likely that carbonates are separated here from sandstones by a thrust plane. In addition, the whole section is poorly exposed, and contacts between different lithologic bodies are usually concealed. Accordingly, it is difficult to interpret relationships between various fragments of the Kusa block.

As is seen in exposure U-2 (Fig. 1), the siliciclastic–dolomitic succession with cherts is below beds with *Tyssagetes* stromatolites, but the direct contact between two units is unexposed, and its character is enigmatic. At any rate, the upper beds with *Tyssagetes* stromatolites are exposed in a series of outcrops along the railway between the Kusa and Zlokazovo stations, and on the right side of the Ai River (exposure U-3) immediately upstream of the railway bridge. Vlasov [13] described a group of *Tyssagetes* forms from the last exposure, and exactly the beds with these stromatolites are regarded in most publications as the basal member

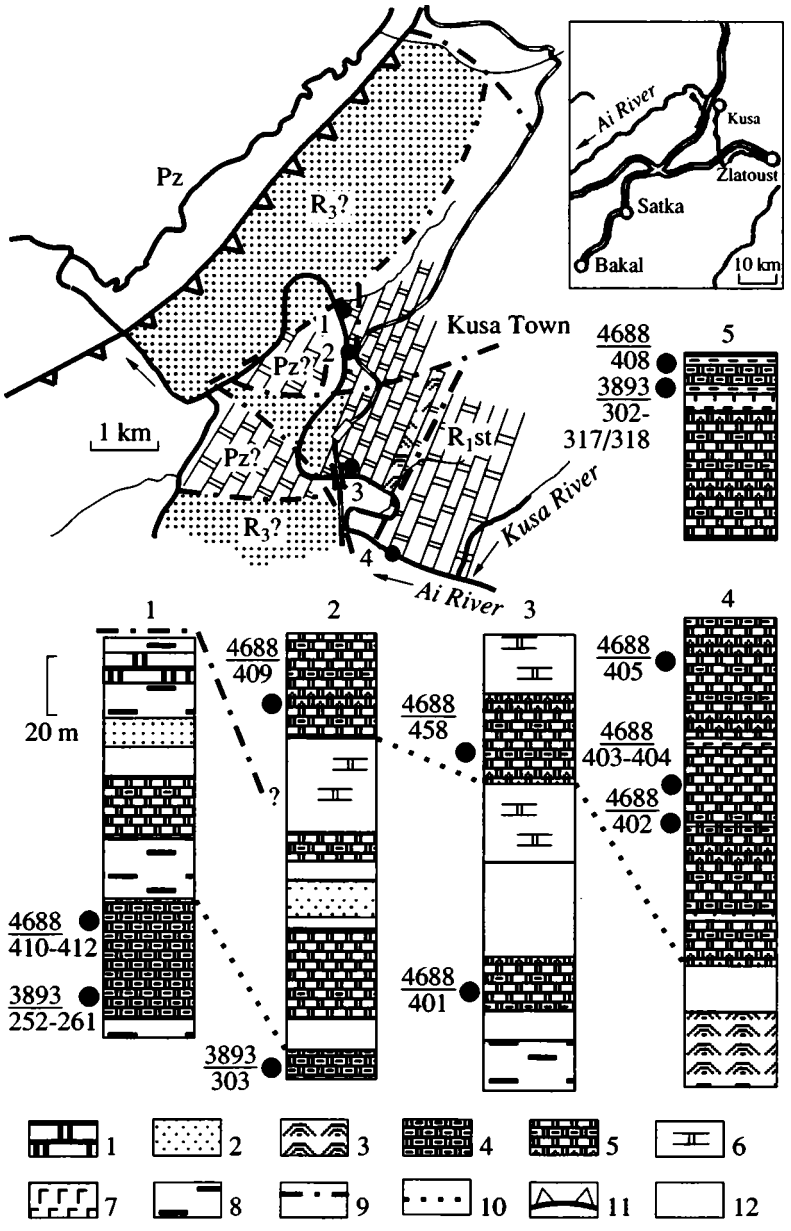


Fig. 1. Structural scheme of Riphean deposits in the Kusa area and inferable stratigraphic relationships between rock successions bearing silicified microfossils (inset map shows geographic position of study area): (1) dolostone; (2) sandstone; (3) columnar stromatolites *Kussiella kussiensis*; (4) dolostone with cherty lenses and interlayers; (5) dolostone with columnar-laminated and laminated silicified stromatolites; (6) poorly exposed intervals with dolostone debris; (7) diabase; (8) shale and claystone; (9) faults observed and inferable; (10) presumable correlation lines for particular members of the Satka Formation; (11) principal thrust fault between the Riphean and overridden Paleozoic rocks; (12) unexposed section intervals (R_1st, Lower Riphean, Satka Formation). Column numbers correspond to section localities numbered in the scheme; exposure no. 5 is situated northeastward of the scheme area.

of the Satka Formation. The rock succession overlying this member and exposed between the railway bridge and the Argus Cliff is undoubtedly stratigraphic, considered as the type section of the lower and upper Kusa subformations. It has been described in detail by Maslov and Krupenin [14], but according to our objectives we describe below only several units distinguished here by Garan' [15].

In the classical section exposed on the right side of the Ai River downstream and upstream of the Kusa factory, the following units are distinguishable in the upper Kusa Subformation (from the base upward): (I) Gray to bluish gray fine-grained dolostones and dolomitic limestones with lenses of flat-pebbled conglomeratic breccia and with interbeds of greenish and gray sericite–chlorite–quartz and sericite–carbonate slates (350 m

thick); (II) Gray to dark gray dolostones grading into sandy varieties and frequently bearing angular to subangular quartz and feldspar grains (150 m thick); (III) Gray to dark gray stromatolitic, predominantly biohermal dolostones with *Kussiella kussiensis* and other stromatolite forms (30–40 m thick); (IV) Red-brown and greenish sericite–quartz and marly slates with limestone and fine-grained sandstone interbeds (15–20 m thick); (V) Light to bluish gray fine-grained dolostones with abundant lenses of flat-pebbled conglomeratic breccia (120 m thick).

The overlying part of the Satka Formation is likely exposed near the town cemetery (exposure U-4) as it is accepted in many publications. Trying to understand relationships between this section and the unit succession described above, we established however that the unit V and likely the unit IV cannot be distinguished in outcrops near the cemetery, and that upper cherty–carbonate beds are faulted here against the lower rocks of unit III with stromatolites *Kussiella kussiensis*. In these outcrops, cherty dolostones enclose *Tyssagetes* stromatolites, which are similar to stromatolites in basal horizons of the Satka Formation. Consequently, there is a great probability that in this case we observe a tectonic slice, in which the above succession is repeated. This is only the assumption that cannot be proved, because two sections under comparison are not identical.

The section part, which is stratigraphically higher, can be examined on the right bank of the Kusa River near the railway to the Kusa factory. Dolostones, the cherty stromatolitic varieties included, are intercalated here with abundant siliciclastic interbeds. Being impregnated by numerous diabase dikes and sills, the rocks are badly altered and recrystallized, deprived therefore of microfossils. Despite the similarity between stromatolites from this and above locality near the cemetery, abundant siliciclastic interbeds in the section under consideration suggest a higher stratigraphic position of exposed rocks.

In the Kusa area, the uppermost rocks of the Satka Formation are exposed on the left side of the Kusa River and in a series of outcrops on both sides of the Ai River. Light gray dolostones with rare claystone interlayers and dark gray cherty lenses are most widespread here. Some of the rocks seem to be phosphatic, although we did not analyze them specially. In the road cut at the right bank of the Ai River (exposure U-5 near the Kusa–Medvedevka road), the above rocks are crowned by the dolostone member with greatly extended (stratal) lentils of phosphatic cherts. At a higher hypsometric level, there is a succession of quartz-sandstone beds formerly attributed to the Mekhovaya Subformation of the Satka Formation [16]. After the detailed examination [17], it was figured out that this succession is in allochthonous position and has no relation to the Satka Formation. Farther along the Ai River between the Kusa Town and Medvedevka Settlement, there are exposed rocks of the Ai Formation. Sections of the lower and upper Kusa subformations are traceable in a narrow belt extending to the Berdyaush Town, where these deposits are intruded by rapakivi bodies.

The lower age limit of the Riphean type succession is defined by U–Pb and Rb–Sr isotopic dates (2000 to 2200 Ma) obtained respectively for zircons and feldspars from gneisses of the Taratash Complex. The K–Ar dating showed that biotite and microcline from metasomatically altered gneisses are 1660–1736 and 1700–1800 Ma old, respectively, whereas the Rb–Sr date for muscovite from a granite vein cutting the altered gneisses is equal to 1690 Ma [18]. Several generations of zircons established in trachybasalts of the lowermost unit of the Ai Formation have been analyzed to determine the age of their host rocks [19]. Despite all the relevant difficulties in age determination, the analytical results have been used to assess the formation and transformation ages of trachybasalt bodies as corresponding to 1635 ± 30 and 620 ± 30 Ma, respectively [17].

The upper chronometric limit of the Burzyan Group is defined by a series of isotopic dates characterizing the emplacement time of hypabyssal gabbro intrusions and Berdyaush rapakivi massif. The rapakivi massif is 1348 ± 13 to 1354 ± 20 Ma old according to results of Rb–Sr isochron and U–Pb zircon dating methods [19]. The concordant dates have been obtained for volcanic rocks of the Mashak Formation, the basal one in the Middle Riphean Yurmata Group. The Rb–Sr date of 1346 ± 41 Ma characterizes whole-rock samples of dacite–rhyolite volcanic series, while the U–Pb dating of zircons from these volcanic rocks yielded the age value of 1350 ± 30 Ma [20]. Thus, it is possible to infer from above data that isotopic ages of lower boundaries of the Burzyan and Yurmata groups are 1650 ± 50 and 1350 ± 50 Ma old, respectively.

COMPOSITION AND SIGNIFICANCE OF MICROBIOTA FROM THE SATKA FORMATION

I.N. Krylov was first to discover by the end of the 1970s the silicified microfossils in rock samples collected by B.M. Keller from the Satka Formation. However, the microfossils had not been described formally, and first publication devoted to the Satka microbiota appeared only in the second half of the 1980s, when additional samples collected by V.N. Sergeev considerably augmented the original collection [21]. Later on, after further sampling of microfossils and taxonomic systematization of formerly identified taxa, Sergeev described the silicified Satka microbiota in his monograph [22]. All the described representatives of that microbiota were found in two localities of the Asha structural-lithologic zone within the Kusa Town area, while cherty rocks sampled in the eastern Kisegan zone turned out to be barren of microfossils.

Somewhat earlier, Yankauskas and Keller detected and described the organic-walled microfossils from core samples of the Satka Formation recovered by boreholes near the Bakal Town in the eastern Kisegan structural-lithologic zone [23]. This microfossil assemblage was comprehensively described by Yankauskas [24] in the monograph dedicated to the Riphean stratotype. In addition, the organic-walled microfossils presumably characterizing the Lower Riphean were detected in core samples of boreholes drilled in the Bashkirian Cis-Urals [25]. Sedimentary successions bearing these microfossils are of a dubious stratigraphic position, and their microfossils are considered below in the course of general biostratigraphic analysis of Early Riphean microbiotas known in the Urals and other regions.

The established microphytological characterization of organic-walled and silicified microfossil assemblages is consistent in general. Both the cherty and shaly rock types yield remains of morphologically simple filamentous and coccoidal microorganisms whose assemblages are similar in principle. Nevertheless, representatives of silicified microbiota have been mostly interpreted as taxa of benthic cyanobacterial communities, e.g., as species of genera *Siphonophycus, Oscillatoriopsis, Palaeolyngbya, Gloeodiniopsis, Eosynechococcus*, and others [26], whereas the organic-walled specimens have been considered almost exclusively as microorganisms of phytoplankton, for instance as representatives of genera *Protosphaeridium, Leiosphaeridia, Symplassosphaeridium, Satka, Micrhystridium*, and others [24]. Veis *et al.*, [27] who continued the research discovered and described the organic-walled microfossils of the Satka assemblage in the western sections of the Kusa Town area. They identified a series of new morphotypes as compared to those described by Yankauskas and Keller, for instance the genera *Nucellosphaeridium* and *Leiotrichoides*, although their work did not change in principle the former characterization of microbiota. The established relationships between the organic-walled and silicified microfossil assemblages seemed normal from the viewpoint of Precambrian microphytology of the terminal 1980s. Later on, it became clear, however, that the Late Riphean silicified microbiotas include abundant microfossils of phytoplankton group, which are of a complex morphology [28]. These complex morphotypes suggested that planktonic forms of similar kind can be found in older microfossil assemblages as well.

During the field seasons of 1998, 1999, and 2000, Sergeev additionally sampled all the microfossil localities in the Kusa Town area. After the careful analysis, new materials showed that benthic cyanobacterial remains of the Satka microbiota indeed coexist with microfossils of phytoplankton group, which have been known before almost exclusively in assemblages of organic-walled remains from siliciclastic deposits. Moreover, a thorough revision of formerly studied thin sections (collections of B.M. Keller and author) resulted in identification of new phytoplanktonic forms, which have been either missed or interpreted as benthic cyanobacterial taxa by original investigation. Thus, it was established that the assemblage of silicified microfossils under examination is sufficiently diverse and consist of 23 species representing 15 genera; five of the species are identified in open nomenclature (Fig. 2).

Studied localities. In the Kusa Town succession exposed on the right bank of the Kusa River, cherts of the Satka Formation have been sampled at the following five localities (Fig. 1): (1) Exposure U-1, the Ulin Kamen Cliff, 3.5 km downstream of the town (samples nos. 3893/252-261 and 4688/410-412); (2) Exposure U-2, 2.5 to 3.0 km downstream of the town (samples nos. 4688/409 and 3893/303); (3) Exposure U-3, shortly upstream of the railway bridge (samples nos. 4688/401 and 4688/458); (4) Exposure U-4 near the town cemetery (samples no. 4688/402-405); and (5) Exposure U-5 in a road cut near exit to the Medvedevka Village (samples nos. 4688/408, 3893/302, 3893/317, and 3893/318).

Composition of microbiota. After taxonomic revision and additional research, it is established that the Satka microbiota consists of the following taxa (Fig. 2): *Palaeolyngbya catenata* Hermann, *Oscillatoriopsis vermiformis* (Schopf), *Filiconstrictosus* sp., *Eomicrocoleus* sp., *Siphonophycus septatum* (Schopf), *Siphonophycus robustum* (Schopf), *Siphonophycus typicum* (Hermann), *Siphonophycus kestron* Schopf, *Siphonophycus solidum* (Golub), *Archaeoellipsoides major* (Golovenoc et Belova), *Eosynechococcus moorei* Hofmann, *Eosynechococcus amadeus* Knoll et Golubic, *Gloeodiniopsis lamellosa* Schopf, *Sphaerophycus medium* Horodyski et Donaldson, *Coniunctiophycus gaoyuzhuangense* Zhang, *Corymbococcus rexii* sp. nov., *Myxococcoides inornata* Schopf, *Myxococcoides* sp., *Leiosphaeridia atava* (Naumova), *Leiosphaeridia crassa* (Naumova), *Pterospermopsimorpha pileiformis* Timofeev, *Granomarginata ?* sp., and of unnamed planktonic morphotypes. The very representative and diverse assemblage is characteristic of the lower Kusa Subformation, whereas the upper one yields only rare specimens of the same microfossils. Microfossils from both subformations are united into a single assemblage named the Satka microbiota. Representatives of microbiota are most abundant and diverse near the base of the lower subformation observable in Exposure U-1, the Ulin Kamen Cliff. Assemblages from other localities represent impoverished versions of this microbiota. It is remarkable as well that some taxa, for instance *Gloeodiniopsis lamellosa* and characteristic new, comparatively large phytoplanktonic forms are found in exposures U-1 (Ulin Kamen Cliff) and U-2 (about 1 km upstream of the latter) only (Fig. 1), whereas taxa from other localities are of a wide stratigraphic range. Microfossils are confined to cherts in microlaminated dolostones or flat-pebbled silicified conglomeratic breccias.

Mat-forming microfossils. Empty sheaths of the genus *Siphonophycus* Schopf, which are dominants of the Satka microbiota, are attributed to five widespread Precambrian species: *S. septatum* (Schopf), *S. robustum* (Schopf), *S. typicum* (Hermann), *S. kestron* Schopf, and *S. solidum* (Golub) with diameter ranges of 1.0–2.0, 2.0–4.0, 4.0–8.0, 8.0–16.0, and 16.0–32.0 μm, respectively (Plate III, figs. 4, 8, 10, 11). These are the mat-forming forms occurring in colonies formed by hundreds of intricately interlacing individuals. The colonies represent either mats with microbial lamination, which are several microns thick, or isometric rounded forms (clasts) ranging in size from 5–10 × 15–25 to 20–30 × 150–200 μm. The taxa have been traditionally and soundly interpreted as empty sheaths of hormogonian cyanobacteria of the LPP (*Lyngbia–Phormidium–Plectonema*) type.

Thin trichomes *Oscillatoriopsis vermiformis* consisting of isometric cells 0.7 to 1.5 μm in diameter and 0.8 to 1.5 μm long (Plate III, fig.6) are also rather abundant components of the microbiota. Like the sheaths *Siphonophycus*, these trichomes are complicatedly interlaced in colonies and represent undoubtedly the mat-forming morphotypes. The trichomes represent most likely the filaments of those oscillatorian cyanobacteria whose empty sheaths are classed with *S. robustum*. Some morphotypes, e.g., trichomes *Gunflintia* aff. *minuta* Barghorn and sheaths *Siphonophycus* sp., which have been described as components of the Satka microbiota [26], are regarded now, after a comparative analysis and thorough revision of the microbiota composition, as synonyms.

Other filamentous forms are represented in the microbiota under consideration by rare monotrichomatous filaments *Palaeolyngbia catenata* (sheaths 20.0 to 25.0 μm in diameter, trichomes 10.0 to 15.0 μm across) and by rather frequent polytrichomatous filaments *Eomicrocoleus* sp. (sheaths 10.0 to 40.0 μm in diameter, trichomes 1.0 to 2.0 μm wide). Filaments *Palaeolyngbia catenata* are remains of cyanobacteria that have dwelt in mats and represent a more completely preserved form of microorganisms, the empty sheaths of which are described as *Siphonophycus kestron*. According to the present-day understanding, the genus *Eomicrocoleus* is a form taxon like the genus *Siphonophycus* [29], and filaments of this taxon may be interpreted as fossil analogues of recent polytrichomatous cyanobacteria of genera *Microcoleus, Hidrocoleum*, or *Schizothrix*. On the other hand, they may represent occasional post-mortal accumulations of cyanobacterial trichomes surrounded by organic matter secondary condensed.

Coccoidal symbionts. Spheroids *Gloeodiniopsis lamellosa* 8.0 to 45.0 μm in diameter exemplify one of dominant taxa of the Satka microbiota. They do not form mats but associate with empty sheaths *Siphonophycus* occurring among the latter. It is remarkable that *G. lamellosa* is found in association with remains of those mats only, which are composed predominantly of *S. robustum, S. typicum*, and, to a lesser extent, of *S. kestron* filaments. Colonies of *S. septatum* and *S. solidum* sheaths are always free of other microfossils, although they may occur as minor components in mats of three aforementioned *Siphonophycus* species. Spheroids of the genus *Gloeodiniopsis* have been convincingly interpreted as remains of chroococcacean cyanobacteria of the *Gloeocapsa* or *Chroococcus* type [30]. Bearing in mind a high polymorphism of microfossils representing the genus *Gloeodiniopsis*, we do not exclude, however, that fossil populations of these cyanobacteria include also the remains of lower, morphologically similar protists. It is noteworthy as well that there is a rare but important preservation form of chroococcacean cyanobacteria: in some cases, there were observed pseudospines of secondary origin on their external envelope, for instance, like those of *Gloeodiniopsis lamellosa* from the Middle Riphean Avzyan Formation [31].

Ellipsoidal cells of cyanobacteria *Eosynechococcus moorei* and *E. amadeus* (greater axes of ellipsoids with size ranges of 1.5–12.5 and 0.5–3.5 μm, respectively) should be attributed to other mat-dwelling microorganisms of the Satka microbiota. These are fossil analogues of present-day cyanobacteria representing the genus *Synechococcus*. The direct observations and principle of uniformitarianism imply that they inhabited mats of hormogonian cyanophytes, the empty sheaths of which are attributed to the genus *Siphonophycus*. Current data of molecular biology [32] show, however, that cyanobacteria of the genus *Synechococcus* represent a heterogeneous group of prokaryotic photosynthetic microorganisms. In view of morphological simplicity of microfossils in question, there is a great probability that they represent remains of both the photosynthetic and heterotrophic microorganisms [34]. It is remarkable as well that the best preserved ellipsoids *Eosynechococcus* occur in those *Siphonophycus* mats of the Satka microbiota, which are most badly decomposed. This leads to assumption that ellipsoids *E. moorei* and *E. amadeus* are the remains of heterotrophic bacteria, which developed in colonies of dead hormogonian cyanobacteria [22].

In mat remains of hormogonian cyanophytes, ellipsoids *Eosynechococcus* occur in association with coccoidal microfossils *Sphaerophycus medium* having one- or two-layer envelopes, diameters of which range from 2.0 to 5.5 μm. In the colonies composed of a few to dozens specimens, spheroids of the last taxon have a loose arrangement, being spaced or slightly touching each other by walls. Microfossils of the genus *Sphaerophycus* are also interpreted as remains of unicellular chroococcacean cyanobacteria, although the alternative interpretation of these morphologically simple remains is also admissible like in the case of microfossils representing the genus *Eosynechococcus*. Coccoidal microfossils *Corymbococcus rexii* represent another colonial taxon whose one- or two-layer spheroids (inner diame-

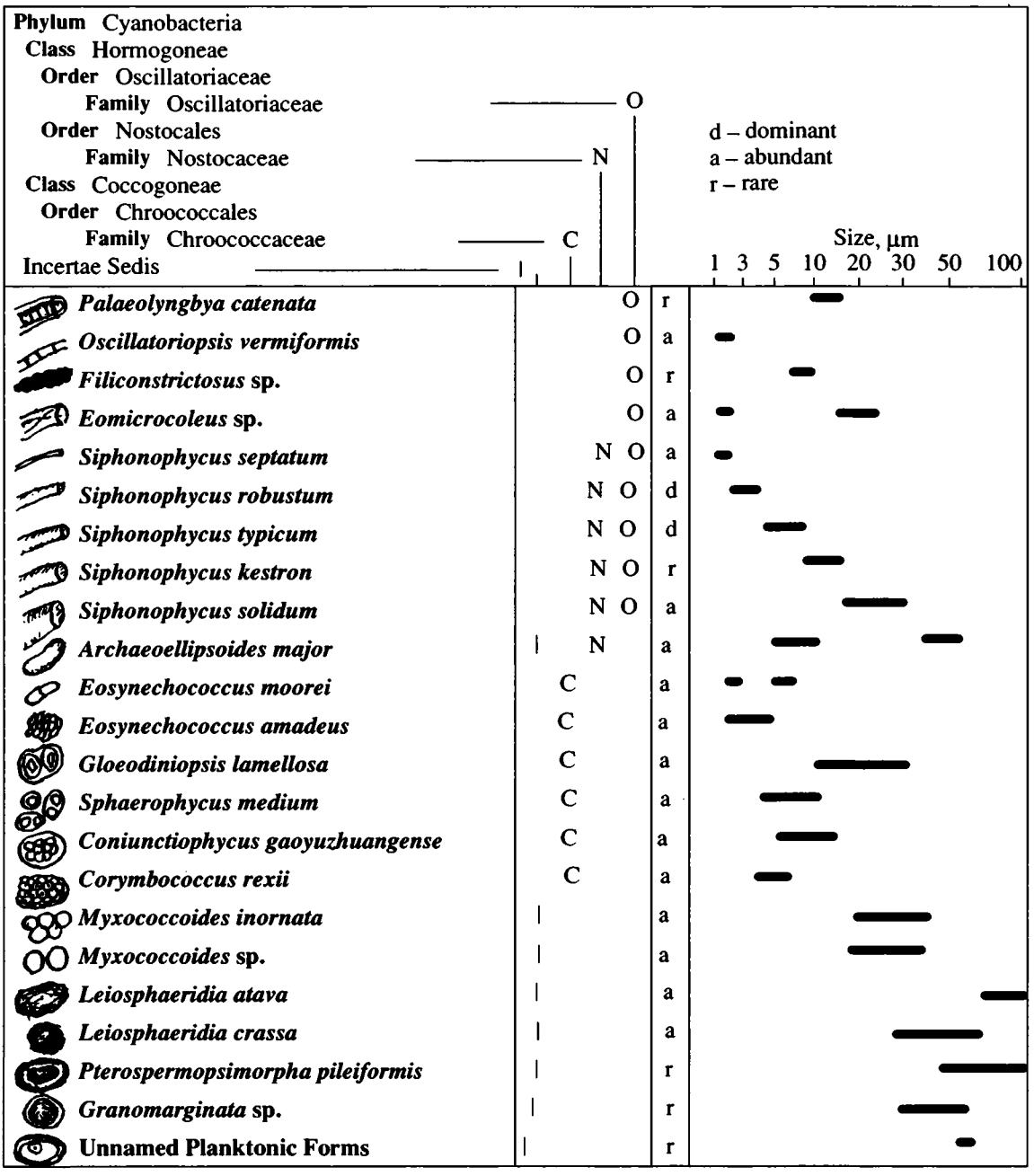


Fig. 2. A synopsis of silicified microfossils from the Satka Formation cherts (size in logarithmic scale).

ters from 2.5 to 5.0 μm and outer diameters from 4.5 to 7.0 μm) occur as loose accumulations surrounded by a common amorphous mass ranging in size from 20.0 × 30.0 to 75.0 × 165.0 μm. Microfossils of the genus *Corymbococcus* have been regarded as fossil analogues of recent chroococcacean cyanobacteria of the genus *Aphanocapsa* [34].

Planktonic forms. Among microfossils presumably representing planktonic components of the Satka microbiota, there are two groups of remains of pro- and eukaryotic microorganisms, as we think. The first group likely includes spherical colonies of small one-layer spheroids (4.0 to 8.0 μm in diameter), which are closely spaced and classed with *Coniunctiophycus gaoyuzhuangense*. These microfossils used to be regarded as ancient analogues of recent chroococcacean cyanobacteria of the genus *Microcystis*, which lead the planktonic life [35]. Ellipsoidal microfossils *Archaeoellipsoides major*, which are 18.5–33.0 × 10.7–17.0 μm in size, should be regarded as remains of planktonic cyanobacteria as well. In distinction from *Synechococcus* forms, they represent spores of *Ana-*

baena-like nostocalean cyanobacteria. This assumption is grounded by the fact that specimens demonstrating cell division have not been detected in the studied fossil population. The microfossils in question are chaotically dispersed within a laminated amorphous mass that presumably represents a decomposed cyanobacterial mat. One-layer spheroids *Myxococcoides* sp. ranging in diameter from 12.0 to 26.0 μm are likely the fossil akinetes, because they occur in a tight association with *Archaeoellipsoides* forms and are similarly dispersed in organic matrix. Species *Myxococcoides inornata* (spheroids 13.0 to 27.0 μm in diameter) from cherts of the Satka Formation are of a less definite origin and may represent either prokaryotic remains, spherical akinetes included, or cells of lower eukaryotic microorganisms.

Remains of presumable eukaryotic phytoplankton are likely represented by a group of large coccoidal microfossils of the Satka microbiota. According to their morphological peculiarities, these microorganisms likely had cell nuclei structurally isolated. The corresponding spheroids of considerable size have coriaceous envelopes, and their colonies are of a relatively complex structure. Nevertheless, being free of real acanthi and processes, the Satka microfossils of this kind are lacking thus the obvious morphological features characteristic of the eukaryotic organization level. It is only a hypothesis that they had cell nuclei, and fossils of the group are considered in this work as Insertae Sedis.

Another peculiar morphotype of the Satka microbiota is *Pterospermopsimorpha pileiformis*. The compact coriaceous envelopes of this taxon range in diameter from 45.0 to 98.0 μm and bear inside dense oval inclusions with greater axis from 30.0 to 85.0 μm long. Spheroids *Pterospermopsimorpha pileiformis* exceed in size the dimension admissible for cells of prokaryotic microorganisms, and their large and dense cyst-like inner inclusions exclude a possibility that the forms characterize empty colonies of prokaryotic microorganisms. Among other presumable remains of unicellular planktonic eukaryotic microorganisms, there are spherical microfossils with thick opaque granular or thin transparent fine-grained walls. Two groups of these microfossils present in microbiota (27.0–68.0 and 130.0 μm in size) are attributed to species *L. crassa* and *L. atava* of the form genus *Leosphaeridia*. Also peculiar are microfossils classed with *Granomarginata ?* sp. These are represented by dark central bodies 30.0 to 35.0 μm in diameter, which are surrounded by less dense envelopes (rimmed by "zones") ranging in diameter from 35.0 to 40.0 μm. Some of them may characterize decomposition stages of other taxa, e.g., of *Pterospermopsimorpha pileiformis* (Plate III, figs. 8 and 9). The others (Plate III, figs. 10 and 11) are probably the slightly altered remains of planktonic microorganisms resembling the two-layer microfossils from younger deposits.

The last form to be mentioned is the multilayer spheroid about 55.0 μm in diameter, which is regarded here as unnamed planktonic morphotype of protists. The single distinguished specimen has a structural element resembling spines (Plate III, fig. 7), but we desist from biologic interpretation and Latin nomenclature in this case, because the material is not representative.

Comparison of microfossils from shaly and cherty–carbonate facies of the Satka Formation. As is mentioned above, microfossil assemblages from shaly [36] and cherty–carbonate facies of the Satka Formation [26] used to be regarded initially as considerably different in composition. Our research of cherty–carbonate facies showed however that benthic forms, which are localized within these facies predominantly in plat-pebbled conglomeratic breccias and laminated fossil mats, occur in association with remains of planktonic microorganisms, which occur inside and outside the mats and their clasts. The change in view on the Satka microbiota composition is determined not only by new materials, but also by changes in the old concept of the 1970s–1980s, which predicted that accumulations of planktonic forms are mostly confined to shales, while benthic cyanobacterial communities are mostly buried in cherts. Accordingly, the initial study of Uralian silicified microfossils was mostly devoted to analysis of benthic cyanobacterial remains, which had been already described in publications, while coexisting planktonic forms were left without attention [26]. At present, it is possible to state that remains of benthos and phytoplankton occur in both preservation forms as organic-walled and mineralized microfossils. A good example is the Satka microbiota as it can be viewed in the light of new data. Microfossils of this microbiota are typical representatives of pre-Late Riphean (Mesoproterozoic) microorganisms.

Lateral distribution and paleoecologic interpretation. Microbiota of the Satka Formation obviously dwelt in shallow-water settings, likely in the upper subtidal and intertidal zones, as it is evident from abundance of conglomeratic breccias and oncolites in deposits bearing microfossils. Nevertheless, the microbiota is lacking remains of entophysalidacean cyanobacteria, the typical representatives of shallow-water arid basins.[1] The facies confinement of microfossils is well detectable in two nearby exposures U-1 (Ulin Kamen Cliff) and U-2, where their assemblage was found close to the Satka Formation base. In both sections, microfossils are confined to flat-pebbled conglomeratic breccias. In addition, planktonic forms present among microfossils and absence of entophysal-

[1] By initial description, *Gloeocapsa*-like colonies from the upper Kusa Subformation were identified as *Eogloeocapsa* sp. [26]. After reexamination of thin sections, it was concluded that cherts of this subformation bear cyanobacterial species *Eoentophysalis dismallakesensis*, but this conclusion has not been proved afterward. The species is identical to characteristic taxon of the Min'yar Formation, and we suspect a mistake by preparation of thin sections.

idacean cyanobacteria are indicative of a shallow-water habitat in the upper subtidal zone or, possibly, of a somewhat deeper proximal shelf above the storm-wave base.

The facies positioning of microfossils at other stratigraphic levels of the Satka Formation is defined less confidently. In most cases, the microfossils are confined to cherty stromatolites of the *Paniscollenia* type or to other nodular or laminated forms (exposures U-2, U-3, and U-4). In these exposures, stromatolites occur in association with the flat-pebbled conglomeratic breccias, which are silicified and bear microfossils. Since microfossil assemblages from all the localities are similar, the entire cherty–carbonate succession studied was likely deposited in the shallow-water subtidal settings.

Biostratigraphic and evolutionary significance of the Satka microbiota. Being represented only by simple filamentous and coccoidal forms, the Satka microbiota looks conservative like many other pre-Late Riphean microbiotas. Size ranges of Satka microfossils are also comparable with those known for concurrent microbiotas. Nevertheless, the Satka microbiota differs in taxonomic composition from most silicified assemblages of the Early Riphean microorganisms. The most characteristic and representative assemblages of the Lower–Middle Riphean silicified microfossils are known from the Kotuikan and Yusmastakh formations of the Anabar Uplift (they are termed further as assemblages of the Kotuikan type). The key feature of these assemblages is a high percentage of two characteristic morphotypes. These are the ellipsoidal forms representing the genus *Archaeoellipsoides*, which are regarded as akinetes of nostocalean and stigonemaecean cyanobacteria, and the short trichomes corresponding to germinating forms of these bacteria [37]. In addition, these assemblages include the rather abundant entophysaledacean cyanobacteria, *Synechococcus* forms, and remains of septate and non-septate filamentous microorganisms of genera *Oscillatoriopsis, Siphonophycus, Circumvaginalis*, and some others [38]. Remains of real eukaryotic microorganisms have not been encountered in silicified microbiotas of the Anabar Uplift, although some of their microfossils, for instance, the problematic, ideally spherical *Myxococcoides grandis* may represent forms with morphologically differentiated nuclei.

Microbiotas of the Kotuikan type are widespread in the Lower and Middle Riphean deposits. They have been found in the Gaoyuzhuang and Wumishan formations of China, which are 1620–1400 Ma old [39], in the Dismal Lakes Group of Canada, which is about 1200 Ma old [40], in the Kyutyungda [41] and Debengda [42] formations of the Olenek Uplift, and in the Kheinjua Formation of India [43]. The main types of microorganisms mentioned above are best represented in the Kotuikan Formation, while in other microbiotas their proportions are variable. At the same time, entophysalidacean cyanobacteria and ellipsoidal akinetes *Archaeoellipsoides* are present in all microbiotas, although the latter are not dominants in microbiota from the Debengda Formation.

Akinetes *Archaeoellipsoides*, large spheroids *Myxococcoides*, and short multicellular trichomes, which are dominants in microbiotas of the Kotuikan type, are also the components of the Satka microbiota, but they are very scarce in this case. On the other hand, the *Siphonophycus* sheaths with incorporated coccoidal microfossils of the genus *Gloeodiniopsis* are dominants of the Satka microbiota but subordinate components of other concurrent assemblages. Besides the empty *Siphonophycus* sheaths, remains of hormogonian mat-forming cyanobacteria are represented in the Satka microbiota by multicellular trichomes and entire filaments attributed to form genera *Palaeolyngbia, Oscillatoriopsis*, and *Gunflintia*. In addition, the Satka microbiota includes other, morphologically simple coccoidal microfossils, which are predominantly represented by chroococcacean cyanobacteria of the genus *Eosynechococcus*.

These distinctive features suggest that silicified microbiotas of the Lower–Middle Riphean are represented by two biofacies of the Kotuikan and Satka type. In distinction from the former, the latter are of a limited distribution in the pre-Late Riphean deposits. It is likely that in the Riphean hypostratotype area of the Uchur–Maya region the Svetlyi microbiota from the Aimchan Group 1300–1100 Ma old is of the Satka type [44]. A similar microbiota is known in the southern Urals from the Revet Subformation of the Avzyan Formation. In this terminal unit of the Middle Riphean stratotype, there were found only sheaths *Siphonophycus* and multilayer spheroids *Gloeodiniopsis lamellosa* [31]. It is remarkable as well that taxonomic distinctions between the Early and Middle Riphean microbiotas of both the Satka and Kotuikan types are insignificant. Accordingly, it is reasonable to consider them as the undifferentiated Early–Middle Riphean (Mesoproterozoic) assemblages of silicified microfossils.

Remains of eukaryotic phytoplankton present in the Satka microbiota enable its correlation with assemblages of the organic-walled microfossils. The silicified microfossils from the Satka Formation are undoubtedly comparable with several organic-walled microbiotas of the Early–Middle Riphean age, for instance, with the Satka and Bakal microbiotas of the Lower Riphean stratotype in the Urals, and with the Omakhta microbiota of the Uchur–Maya region in Siberia [45]. The common feature of all microbiotas under comparison is predominance of morphologically simple filamentous and coccoidal forms, which are remains of prokaryotic and unicellular eukaryotic microorganisms from genera *Leiosphaeridia, Pterospermopsimorpha, Myxococcoides, Siphonophycus, Oscillatoriopsis*, and some others. Large morphotypes of complex morphology have not been encountered in the Satka, Bakal, and Omakhta formations.

The Lower Riphean organic-walled microbiotas of the other type are known from marine facies of the Ust-Il'ya and Kotuikan formations of the Anabar Uplift [46] and from the Kaltasy Group of the Cis-Urals, the analogue of the Burzyan Group [47]. Besides the small, morphologically simple filamentous and coccoidal microfossils of genera *Leiosphaeridia, Ostiana (= Coniunctiophycus?), Sphaerocongregus*, sheaths *Siphonophycus* and *Rectia*, akinetes *Brevitrichoides (= Archaeoellipsoides)*, wide trichomes *Botuobia*, and other forms known from organic-walled microbiotas mentioned above, microbiotas of the type under consideration include large sphaeromorphids of the genus *Chuaria* up to 1 mm in diameter. Branching thallomes of the *Ulophyton* type and some other forms of complex morphology have been identified in the Kotuikan and Ust-Il'ya microbiotas [48], but their biologic interpretation is ambiguous in view of possible post-mortal superimposing of originally non-branching filaments [49]. At the same time, acanthomorphic and herkomorphic acritarchs or other Precambrian microfossils of complex morphology, which are characteristic of the Upper Riphean, have not been detected in above microbiotas.

Available data of geochronology, chemostratigraphy, and stromatolitic biostratigraphy soundly suggest the Early or Early–Middle Riphean age of sedimentary units bearing microbiotas of the last type, but investigators of these microbiotas doubted their pre-Late Riphean position in the biostratigraphic succession. In accord with hypothesis predicting that dimensions and morphological complexity of microorganisms should increase with time, they place the host deposits of microbiotas at the Upper Riphean level and consider them as analogues of the Kerpyl Horizon of the Uchur–Maya region [50]. In the Uchur–Maya and Turukhansk regions of Siberia, abundant and diverse microbiotas occur in the succession beginning from the Totta and Bezymyannyi formations only [51]. In addition to morphotypes typical of the Lower Riphean Omakhta microbiota, microfossil assemblages from these formations include large spherical *Chuaria* (up to 1100 μm across), filamentous *Asperatofilum, Taenitrichoides, Rectia, Rugosoopsis, Trachytrichoides*, branching thallomes *Ulophyton* and *Majaphyton*, and some other forms similar to organic-walled microfossils of the Kotuikan Formation.

A special case is microbiota from the Roper Group of Australia, which is 1.5 Ga old according to results of U–Pb zircon dating [52]. Shales of the group yield the oldest remains, as they seem, of microorganisms with true spines and processes, which are attributed to the genus *Tappania* and occur in association with complex *Valeria, Dictiosphaera*, and *Satka* morphotypes [53] and with large filaments and spheroids up to 100 and 500 μm in diameter, respectively [54]. Analogous assemblages were discovered in the Baicaoping and Beidajiang formations of the Ruyang Group in China [55], which are of a problematic age ranging from the Early to terminal Middle Riphean. According to U–Pb zircon dating, granite intrusions in the Ruyang Group are 999 Ma old [56], and $\delta^{13}C_{carb}$ values in the lower portion of the group are close to zero ($0 \pm 1‰$), whereas in upper horizons they are as high as +4‰, i.e., characteristic of C-isotope global excursion recorded in deposits about 1250 Ma old [57]. Similar acanthomorphic forms are unknown in the pre-Late Riphean microbiotas of the Urals and Siberia. To explain this fact, Javaux *et al.*, [53] suggested that acanthomorphic acritarchs of the genus *Tappania* dwelt in the distal shelf settings, whereas Uralian and Siberian assemblages inhabited shallower zones. In the Katanga depression of Siberian platform, the Kama Formation that is older than 1350 Ma according to the isotopic-geochronological dates yielded the representative Kerpyl–Lakhanda assemblage of microfossils, among which there are forms similar to those from the Roper Group, in particular, that assemblage includes acritarchs of the genus *Tappania* (Nagovitsyn, 2001).

If the Roper and Ruyang groups are dated correctly, the appearance of acanthomorphic acritarchs may be confined to the boundary separating the Riphean and pre-Riphean strata. Acanthomorphic acritarchs from the Roper and Ruyang groups differ in taxonomic composition from the Late Riphean acanthomorphic microfossils, among which representatives of genera *Tappania* and *Shuiyousphaeridium* have not been detected so far. Nevertheless, other forms of complex morphology, for instance, *Valeria* and *Dictiosphaera* species are characteristic of the Late Riphean microbiotas. Therefore, it would be premature to conclude that these microbiotas are of the Early Riphean age, in distinction from biotas of the Anabar Uplift. The incorrect decision may be connected with the misleading isotopic-geochronological information or with the erroneous interregional correlation and cryptic hiatuses in the rock succession.

Thus, distinctions between the Satka and some other Lower Riphean microbiotas are determined most likely by the facies factor. On the other hand, the Satka microbiota is obviously similar to the other silicified and organic-walled Lower–Middle Riphean microbiotas. Accordingly, it is possible to think that two or more concurrent types of microfossil assemblages existed during the Early Riphean, and this was probably a consequence of facies peculiarities in distribution of microorganisms and possible paleogeographic zoning of that time.

CONCLUSION

The research carried out augments the database on microfossils from the Lower Riphean stratotype and on general taxonomic composition of microbiotas from cherty rocks. Found in the latter are abundant and diverse microphytological remains of microorganisms formerly known in a form of organic-walled microfossils from shaly rocks only, and this is a most important

result of the work. In addition, we revised the taxonomic composition of described remains representing benthic cyanobacteria and identified several new forms of the latter. The integral taxonomic composition of microbiota from the Satka Formation shows that it is comparable with many coeval assemblages of organic-walled and silicified microfossils, being simultaneously different from most typical microbiotas of the Lower–Middle Riphean time, the dominant forms of which are ellipsoidal akinetes and short multicellular trichomes. At the same time, new data accentuated distinctions between the Satka and some characteristic Lower Riphean organic-walled microbiotas. The established distinctions (absence of such morphotypes as large sphaeromorphic and acanthomorphic acritarchs, thallomes of considerable size, and some others) may be a consequence of the facies or paleogeographic control. Thus, the new data on microbiota from the Satka Formation validate the general concept of vertical and lateral distribution of Precambrian microfossils and contribute much to current understanding of relationships between the silicified and organic-walled microfossil taxa.

DESCRIPTION OF MICROFOSSILS

Kingdom Eubacteria Woese and Fox, 1977

Phylum Cyanobacteria Stanier *et al.*, 1978

Class Coccogoneae Thuret, 1875

Order Chroococcales Wettstein, 1924

Family Chroococcaceae Nägeli, 1849

Genus *Coniunctiophycus* Zhang, 1981

Type species: *Coniunctiophycus gaoyuzhuangense* Zhang, 1981

Coniunctiophycus gaoyuzhuangense Zhang, 1981

Plate I, figs. 3, 4, 7

Coniunctiophycus gaoyuzhuangense Zhang 1981 [58], p. 499, Plate 5, fig. 3; Sergeev *et al.*, 1995 [59], pp. 25–26, fig. 12.5

Eomicrocystis irregularis Golovenoc et Belova, 1984 [60], p. 29, Plate II, figs. 16, 17; Yankauskas *et al.*, 1989 [61], p. 91, Plate XIX, fig.16; Yakshin, 1991 [62], p. 22, Plate VIII, fig. 4.

Eomicrocystis elegans Golovenoc et Belova, 1984 [60], p. 29–30, Plate II, figs. 12–15; Yankauskas *et al.*, 1989 [61], p. 91, Plate XIX, figs. 9; Yakshin, 1991 [62], pp. 22–23, Plate VIII, fig. 3

Holotype: paleontological collection of the Beijing University, thin section no. BGP 7803, point 3; Lower Riphean, Gaoyuzhuang Formation; China.

Description. Spheroids with one-layer envelopes, a few or dozens of which are clustered in spherical colonies. The colonies form in turn the aggregates consisting of several primary spherical structures. Spheroids are 4.0 to 8.0 μm in diameter, and colonies range in size from 10.0 to 20.0 μm across.

Comparison. *C. gaoyuzhuangense* is greater in size than *C. conglobatum.*

Distribution: Lower Riphean, Satka Formation, Southern Urals; Lower–Middle Riphean, Gaoyuzhuang Formation, China; Kotuikan and Yusmastakh formations, Anabar Uplift, Siberia.

Material: nearly a hundred of colonies.

Genus *Corymbococcus* Awramik et Barghoorn, 1977

Type species: *Corymbococcus hodgkissii* Awramik et Barghoorn, 1977

Corymbococcus rexii sp. nov.

Plate I, figs. 14, 15

Holotype: paleontological collection of the Geological Institute RAS, no. 4688, specimen no. GIN 717, thin section no. 2-79, outcrop no. U-1.

Etymology from Latin rex, the king.

Description. One- or two-layer spheroids clustered in colonies of a few to hundreds specimens, which are either spaced from each other or osculating via walls inside the obducing amorphous mass. The inner envelope layer of spheroids is translucent or almost opaque sometimes; it is coarse-grained and as thick as 1.0 μm or less. The outer transparent layer is fine-grained, less than 0.5 μm thick. External diameters of spheroids range from 4.5 to 7.0 μm, and the internal ones are 2.5 to 5.0 μm large. Colonies vary in dimensions from 20.0 × 30.0 to 75.0 × 165.0 μm.

Comparison. Spheroids *C. rexii* are lesser in size than spheroids *C. hodgkissii.*

Remarks. Genus *Corymbococcus* represents multilayer colonial spheroids, which are interpreted as remains of chroococcacean cyanobacteria comparable with present-day species of the genus *Aphanocapsa* [34]. Later on, Nyberg and Schopf [63] described genus *Eoaphanocapsa* and regarded it as a possible analogue of *Aphanocapsa.* It is possible that both taxa are synonyms, but their revision is out the scope of our work, because the material necessary for the comparative analysis is too large.

Material: several dozens of colonies.

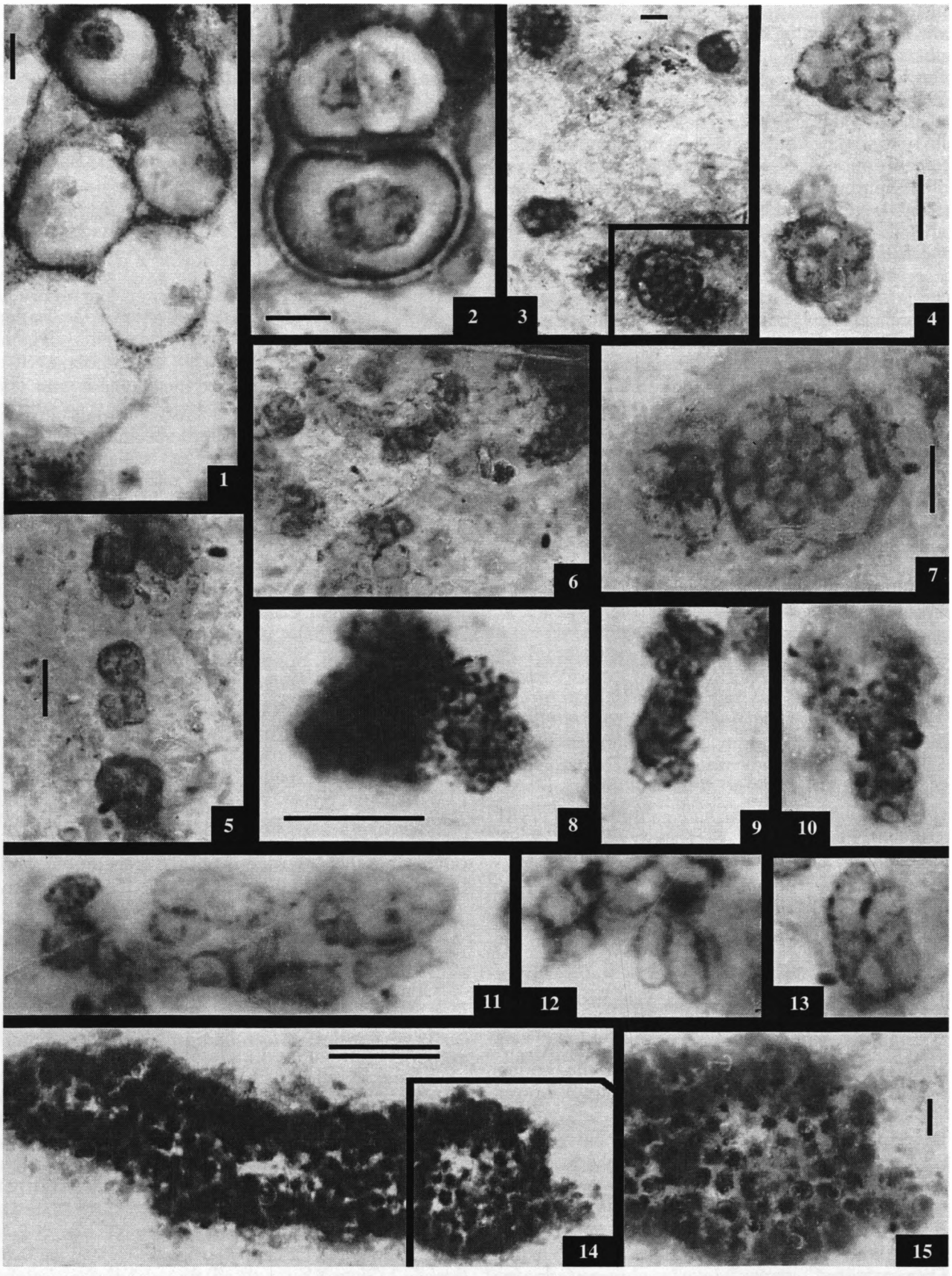
Genus *Gloeodiniopsis* Schopf emend. Knoll et Golubic, 1979

Type species: *Gloeodiniopsis lamellosa* Schopf, 1968

Gloeodiniopsis lamellosa Schopf, emend. Knoll et Golubic, emend. Sergeev,1992

Plate I, figs. 1–2

Gloeodiniopsis lamellosa Schopf 1968 [64], p. 684, Plate 84, fig. 2; Knoll and Golubic, 1979 [65], p. 147, figs. 6, 7; Sergeev, 1992a [22], pp. 74–76, Plate IX, figs. 4–9, Plate X, figs. 1, 3, 4, Plate XI, figs. 1, 3–8, Plate XII, figs. 1, 2, Plate XIII, figs. 1–3, 7–9, Plate XIV, figs. 1–7, Plate XV, figs. 1–7, Plate XVII, figs. 7a –7d; Sergeev *et al.*, 1997 [66], pp. 216–219, fig. 7, 8A–8H.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Gloeodiniopsis uralicus: Krylov and Sergeev, 1986 [21], p. 103, Plate III, figs. 1–7; Yankauskas *et al.*, 1989 [61], p. 94, Plate XXII, figs. 2, 7, 8, 13; Sergeev, 1992 [22], pp. 77–78, Plate I, figs. 1–6; Plate II, figs. 1–3, 5–7; Plate III, figs. 1–4; Plate 4, figs. 1–2 (for a complete synonymy, see [66]).

Holotype: paleobotanical collection of the Harvard University, no. 58502, thin section Bit/Spr 6-3, coordinates 19.1 × 96.6; Late Riphean, Bitter Springs Formation, Australia.

Description. Spheroids and ellipsoids with one- or two-layer to multilayer envelopes, which are composed of more or less similar layers of identical types. An opaque spherical inclusion is seen attached to the internal side of central envelope layer. Spheroids and ellipsoids occur as individual specimens, more commonly as colonies of several to thousands specimens, which are ether spaced or touching each other by lateral walls. In colonies, there distinguishable monads, dyads, and tetrads (cross or tetrahedral) of spheroids, and some spheroid groups have an envelope. Spheroids range in diameter from 8.0 to 45.0 μm, and internal inclusions are from 2.0 to 5.0 μm across (D_{av} = 3.5 μm, N = 32).

Comparison. Spheroids *Gloeodiniopsis uralicus* have been originally described as considerably exceeding in size the largest specimens of *Gloeodiniopsis lamellosa* known at that time. The subsequent revision of *Gloeodiniopsis lamellosa* type population from the Late Riphean Bitter Springs Formation of Australia showed however that real dimensions of these chroococcacean cyanobacteria are much greater than those reported earlier [67]. According to the latest description [66], spheroids *Gloeodiniopsis lamellosa* range in diameter from 8 to 42 μm, thus being indistinguishable in morphometric parameters from *G. uralicus*.

Remarks. In many works, similar spheroids with dark inner inclusions are attributed to other taxa, for instance to the species *Glenobotrydion majorinum* [63]. In this case, dark inclusions in spheroids have been compared with pyrenoids of green algae [63], and this is not deprived of legitimacy [68]. Nevertheless, the inclusions most likely represent the collapsed protoplast of prokaryotic cells, because presence or absence of identical dark inclusions in spheroids is characteristic of fossil remains of true prokaryotic entophysalidacean algae from the Belcher Supergroup of North America [69]. Accordingly, spheroids *Gloeodiniopsis lamellosa* with dark inner inclusions are interpreted by the authors as remains of prokaryotic cells with protoplast collapsed inside [70].

Distribution: the species is widespread in the Riphean silicified microbiotas.

Material: hundreds of specimens.

Class Hormogoneae Thuret, 1875

Order Oscillatoriales Elenkin, 1949

Family Oscillatoriaceae Dumortier ex Kirchner, 1898

Genus *Palaeolyngbya* Schopf, emend. Butterfield, 1994

Type species: *Palaeolyngbya barghoorniana* Schopf, 1968

Palaeolyngbya catenata Hermann, emend. Butterfield, 1994

Plate II, figs. 1–3

Palaeolyngbya catenata Hermann, 1974, pp. 8–9, Plate 6, figs. 5; Butterfield *et al.*, 1994 [71], p. 61, figs. 25 F–25G

Palaeolyngbya maxima Zhang Y., 1981 [58], p. 495, Plate 2, figs. 4, 6, 7 *Palaeolyngbya* cf. *maxima* Sergeev, 1992a [22], pp. 88–89, Plate V, figs. 4, 5

Oscillatoriopsis sp.: Krylov and Sergeev, 1986, p. 100 [21], Plate 1, figs. 3, 4; Yankauskas *et al.*, 1989 [61], Plate XXII, fig. 3; Sergeev, 1992a [22], 88–89, Plate V, fig. 6

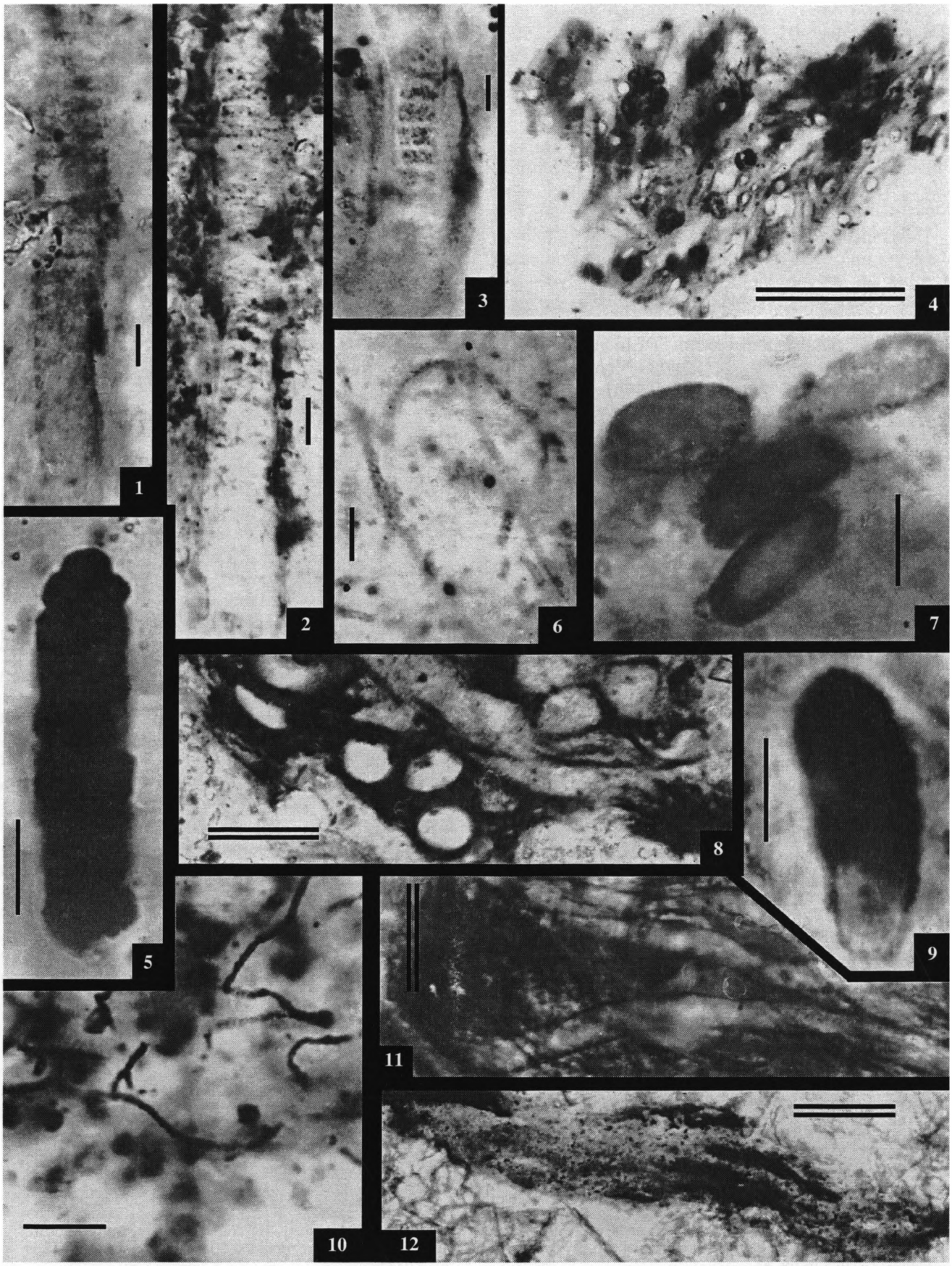
Holotype: paleontological collection of the Institute of Precambrian Geology and Geochronology, St. Petersburg, Russia, preparation no. 49/23; Upper Riphean, Miroedikha Formation, Turukhansk Uplift, Siberia.

Description. Uniform nonbranching trichomes surrounded by a distinct sheath of one- or multilayer structure. Each trichome consists of discoid-cylindrical cells 10.0 to 15.0 μm wide and 2.0 to 5.0 μm high. Crosswalls are distinct, but trichome cells are spaced from each other. Side and cross walls are transparent, medium-grained, less than 0.5 μm thick. Filaments are 20.0 to 25.0 μm in diameter. Trichomes *Palaeolyngbya*

Plate I. Coccoidal microfossils from the Satka Formation.

(1, 2) *Gloeodiniopsis lamellose* Schopf: (1) specimen no. GIN 1, thin section no. 10-80, point 1, (2) specimen no. GIN 3, thin section no. 130-82, point 10; (3, 4, 7—rectangular area in fig. 3) *Coniunctiophycus gaoyuzhuangense* Zhang: (3, 7) specimen no. GIN 712, (4) specimen no. GIN 714, thin section no. 855-01, point 10; (5, 6) *Sphaerophycus medium* Horodyski et Donaldson: (5) specimen no. GIN 715, (6) specimen no. GIN 716, thin section no. 875-01, point 6; (8–10) *Eosynechococcus amadeus* Knoll et Golubic: (8) specimen no. GIN 27, thin section no. 185-82, point 14, (9) specimen no. GIN 28, thin section no. 185-82, point 14, (10) specimen no. GIN 725, thin section no. 185-82, point 1; (11–13) *Eosynechococcus moorei* Hofmann: (11) specimen no. GIN 25, thin section no. 125-82, point 13, (12) specimen no. GIN 24, thin section no. 125-82, point 11, (13) specimen no. GIN 26, thin section no. 125-82, point 13; (14, 15—rectangular area in fig. 14) *Corymbococcus rexii* sp. nov. specimen no. GIN 717, thin section 2-79, point 6 (holotype).

Specimens 1, 24–27, 717, and 725 are from exposure no. U-1, the Ulin Kamen Cliff; specimens 712 and 714–716 are from exposure U-4, the Ai River right bank downstream of the city cemetery; scale bars: ordinary 10 μm, doubled 50 μm (scale of figs. 9–13 is identical to that of fig. 8).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catenata occur as isolated specimens or in colonies of *Siphonophycus* sheaths.

Comparison. In distinction from other *Palaeolyngbya* forms, filaments of *P. catenata* range in diameter from 10 to 30 μm ([71], p. 61).

Remarks. Some trichomes *Palaeolyngbya catenata* have been originally described as *Oscillatoriopsis* sp. [26], because it was suggested [72] that species of the genus *Oscillatoriopsis* may have a sheath less than 1 μm thick. After the subsequent revision of filamentous forms [71], it was decided that the genus *Oscillatoriopsis* includes only morphotypes deprived of sheaths, while all other forms having sheaths of any thickness were attributed to the genus *Palaeolyngbya*, and we include in the latter all the trichomes of Satka microbiota, which are surrounded by sheaths.

Distribution: the species is widespread in Riphean and Vendian deposits.

Material: several, variably preserved specimens.

Genus *Oscillatoriopsis* Schopf, emend Butterfield, 1994

Type species: *Oscillatoriopsis obtusa* Schopf, 1968

Oscillatoriopsis vermiformis (Schopf) comb. Butterfield, 1994

Plate II, fig. 6

Oscillatoriopsis vermiformis Butterfield, 1994, pp. 57–58

Gunflintia aff. *minuta*, Krylov and Sergeev, 1986 [21], p. 97, Plate I, figs. 7; Sergeev, 1992a [22], pp. 97–98, Plate V, fig. 3

Holotype: paleobotanical collection of the Harvard University no. 58467, thin section Bit/Spr 6-7, coordinates 19.4 × 107.4; Late Riphean, Formation Bitter Springs, Australia.

Description. Uniform nonbranching trichomes lacking outer sheaths and occurring as "loose" accumulations of spaced specimens, terminal cells of which have not been preserved. Medial cells are pill-shaped or cylindrical, from 0.7 to 1.5 μm wide and from 0.8 to 1.8 μm long; the width/length ratios range from 1.2 to 1.5. The maximum trichome length is 100.0–150.0 μm (incompletely preserved specimens). Some slightly larger cells have negligible swellings. Cross walls are either distinct or absent; they and side walls are of a single type: translucent, fine-grained, and less than 0.5 μm thick.

Comparison. Dimensions of medial cells differ *O. vermiformis* from other *Oscillatoriopsis* species.

Remarks. The species was originally attributed to the genus *Gunflintia*, because some cells of trichomes are larger than others [21]. The careful comparison with the type population of *G. minuta* showed however that trichomes from the Satka Formation cannot be attributed to that genus, because the enlarged cells retain the relic primary dimensions, whereas the others are contracted in response to selective post-mortal wrinkling of trichomes. The studied specimens are attributed therefore to the genus *Oscillatoriopsis*. Trichomes *O. vermiformis* represent most likely those fossil remains of hormogonian cyanobacteria, the empty sheaths of which are described as *Siphonophycus robustum*.

Distribution: the species is widespread in Proterozoic assemblages of silicified and organic-walled microfossils.

Material: hundreds of specimens.

Filiconstrictosus sp.

Plate II, fig. 5

Description. Solitary nonbranching trichomes lacking external sheaths; their terminal cells are hemispherical in shape. Dolioform medial cells are necked near septae. They are 8.0 to 8.5 μm wide and 2.5 to 3.0 μm long; the width/length ratios range from 2.5 to 3. Terminal cells are as wide and long as 4.5 and 2.5 μm, respectively. The maximum length of trichomes is up to 42.0 μm. Side and cross walls are opaque.

Remarks. The species is very close to *Filiconstrictosus majusculus* Schopf et Blacic, in distinction from which its medial cells are larger in diameter. The studied specimens are described as *Filiconstrictosus* sp. because of a limited material that is preserved not very well.

Material: several variable preserved specimens.

Eomicrocoleus sp.

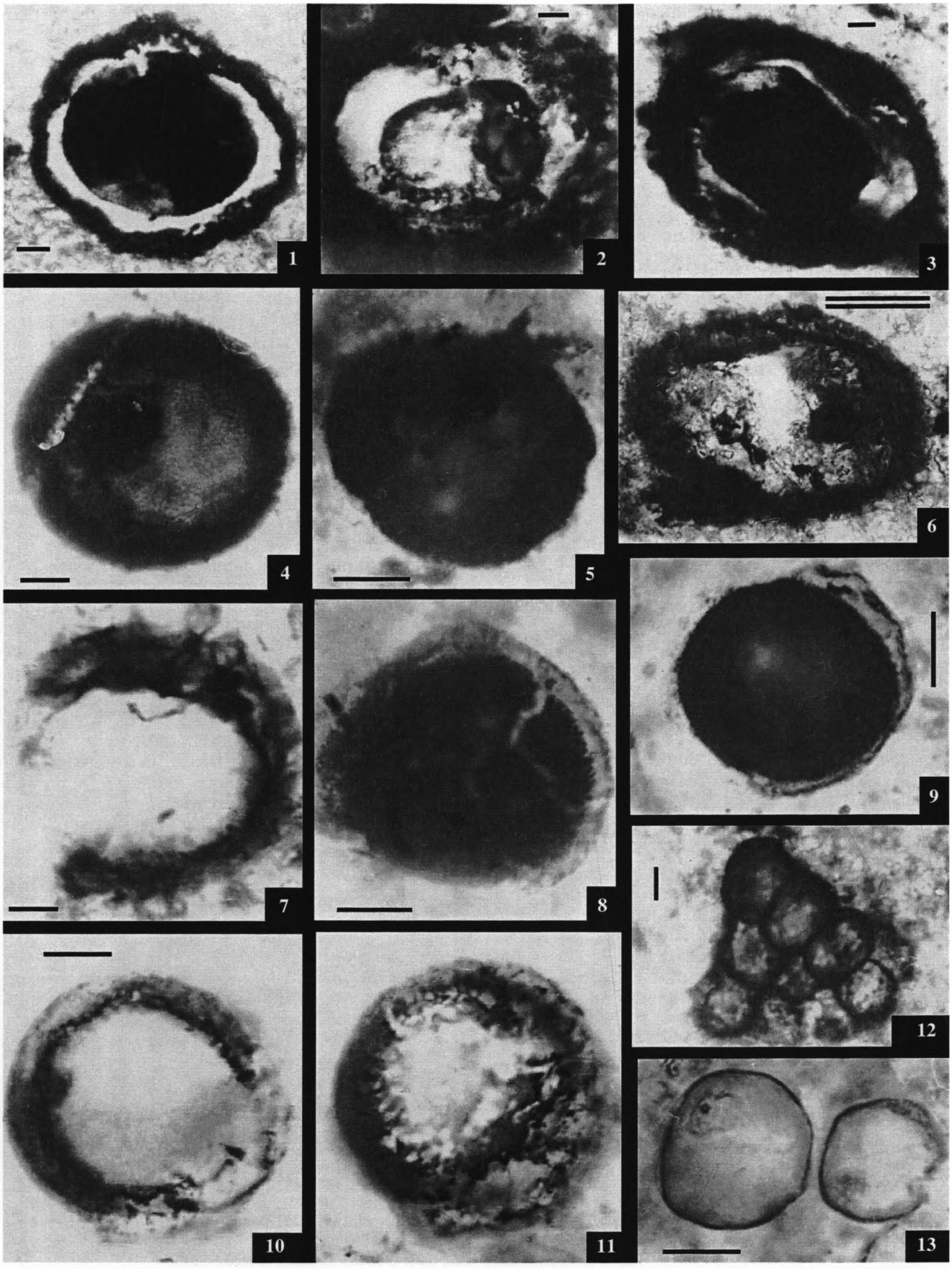
Plate II, fig. 12

Description. Trichomes in parallel arrangement, which are surrounded or not by a common sheath; trichomes are poorly preserved and presumably experi-

Plate II. Filamentous microfossils from the Satka Formation.

(1–3) *Palaeolyngbya catenata* Hermann: (1) specimen no. GIN 35, thin section 120-82, point 3, (2) specimen no. GIN 34, thin section 327-84, point 8, (3) specimen no. GIN 33, thin section 337-84, point 6; (4) *Siphonophycus robustum* (Schopf) and *Siphonophycus typicum* (Hermann), specimen no. GIN 30, thin section 337-84, point 3; (5) *Filiconstrictosus* sp., specimen no. GIN 718, thin section 861-01, point 3; (6) *Oscillatoriopsis vermiformis* (Schopf), specimen no. GIN 32, thin section 182-82, point 32; (7, 9) *Archaeoellipsoides major* (Golovenoc et Belova): (7) specimen no. GIN 719, thin section 872-01, point 17, (9) specimen no. GIN 720, thin section 5-80, point 9; (8) *Siphonophycus solidum* (Golub), specimen no. GIN 721, thin section 866-01, point 10; (10) *Siphonophycus septatum* (Schopf), specimen no. GIN 722, thin section 861-01, point 19; (11) *Siphonophycus kestron* Schopf, specimen no. GIN 723, thin section 865-01, point 21; (12) *Eomicrocoleus* sp., specimen no. GIN 724, thin section 855-01, point 6.

All specimens are from exposure no. U-1, the Ulin Kamen Cliff; scale bars: ordinary 10 μm, doubled 50 μm.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ences the post-mortal compression. Their cross walls are almost indistinguishable, and side walls are coarse-grained. Trichomes are 1.0 to 2.0 μm in diameter. Surrounding sheaths, when they present, range in diameter from 10.0 to 40.0 μm, and their coarse-grained walls are about 1 μm thick.

Remarks. In morphometric parameters, *Eomicrocoleus* sp. from the Satka Formation is practically identical to *E. crassus*, and the form is defined in open nomenclature because of a poor preservation state of studied specimens.

Material: about ten poorly preserved filaments.

Order Nostocalles Geitler, 1925

Family Nostocaceae Kützing, 1843

Archaeoellipsoides major (Golovenoc et Belova) emend. Sergeev et Knoll, 1995

Plate II, figs. 7, 9

Archaeoellipsoides major, Sergeev *et al.*, 1995 [59], p. 31, figs. 10.6–10.8, 10.12–10.15; 11.4–11.8, 13.8; *Eosynechococcus major* Golovenoc et Belova, Golovenok and Belova, 1984 [60], p. 27, Plate II, fig. 2; *Archaeoellipsoides grandis* (partim) Horodyski and Donaldson, 1980 [40], pp. 154–156, figs. 16D-16F (for a complete synonymy, see [59]).

Holotype: collection of the All-Russia Research Institute of Geology, St. Petersburg, Russia, thin section 463-g; Lower Riphean, Kotuikan Formation, Anabar Uplift, Siberia.

Description. One-layer aseptate solitary ellipsoids with rounded endings; their outer surface and sidewalls are translucent, medium- to coarse-grained. The sidewalls are less than 1.0 μm thick. Ellipsoids are 18.5 to 33.0 μm long and 10.0 to 17.0 μm long; the length/width ratios range from 1.8 to 3.0.

Comparison. The described taxon differs in dimensions from other *Archaeoellipsoides* species (see [59]).

Distribution: Middle–Lower Riphean: Dismal Lake and Uluksan formations, Canada; Wumishan Formation, China; Kotuikan and Yusmastakh formations, Anabar Uplift; Satka Formation, Southern Urals.

Material: a few dozens of well-preserved specimens.

Insertae sedis

Genus *Myxococcoides* Schopf, 1968

Type species: *Myxococcoides minor* Schopf, 1968

Myxococcoides inornata Schopf, 1968

Plate III, fig. 12

Myxococcoides inornata Schopf, 1968 [64], pp. 676–677, Plate 84, fig. 7

Holotype: paleobotanical collection of the Harvard University no. 58507, thin section Bit/Spr 6-5, coordinates 27.0 × 93.4; late Riphean, Bitter Springs Formation, Australia.

Description. Spheroids with one-layer envelope; they occur in monads, dyads, triads, and tetrads (cross and tetrahedral) or in colonies of several (up to ten) specimens, which are in a tight contact with each other. The envelopes are usually translucent and spherical. The outer surface and sidewalls of envelopes are medium-grained, and the latter are about 1.0 μm thick. Spheroids range in diameter from 13.0 to 27.0 μm.

Comparison. The species is different in dimensions from other *Myxococcoides* forms.

Remarks. Over thirty species frequently having the overlapping morphometric parameters are described at present as representatives of the genus *Myxococcoides* that is a taxon of heterogeneous microfossils. The species are of a simple morphology, and it is difficult to discriminate them from each other. *M. inornata* is different in dimensions from many other *Myxococcoides* species.

Distribution: the species is widespread in Proterozoic assemblages of microfossils.

Material: several dozens of specimens.

Myxococcoides sp.

Plate III, fig. 13

Description. Spheroids with one-layer envelope, which occur as solitary specimens or as dyads, triads, and cross tetrads; their envelope is translucent. The outer surface and walls are medium-grained, the latter being about 1.0 μm thick. Spheroids range in diameter from 10.0 to 25.0 μm.

Remarks. It is likely that some spheroids *Myxococcoides* sp. are akinetes of cyanobacteria, because they occur in association with ellipsoids *Archaeoellipsoides major*.

Plate III. Remains of phytoplanktonic microorganisms from the Satka Formation.

(1–3, 9) *Pterospermopsimorpha pileiformis* Timofeev: (1) specimen no. GIN 701, thin section 6-80, point 25, (2) specimen no. GIN 702, thin section 6-80, point 25, (3) specimen no. GIN 703, thin section 878-01, point 7, (9) specimen no. GIN 704, thin section 867-01, point 25; (4, 5) *Leiosphaeridia crassa* (Naumova): (4) specimen no. GIN 705, thin section 865-01, point 21; (5) specimen no. GIN 706, thin section 851-01, point 12; (6) *Leiosphaeridia atava* (Naumova), specimen no. GIN 707, thin section 1-79, point 19; (7) unnamed planktonic form, specimen no. GIN 708, thin section 851-01, point 2; (8, 10, 11) *Granomarginata ?* sp.: (8) specimen no. GIN 709, thin section 867-01, point 14; (10, 11) specimen no. GIN 8, thin section 185-82, point 17; (12) *Myxococcoides inornata* Schopf, specimen no. GIN 710, thin section 867-01, point 18; (13) *Myxococcoides* sp., specimen no. GIN 711, thin section 2-79, point 20.

All specimens are from exposure no. U-1, the Ulin Kamen Cliff; scale bars: ordinary 10 μm, doubled 50 μm.

Material: several dozens specimens in a moderate preservation state.

Genus *Leiosphaeridia* Eisenack, 1958

Type species: *Leiosphaeridia baltica* Eisenack, 1958.

Leiosphaeridia atava (Naumova) emend. Jankauskas, 1989

Plate III, fig. 6

Leiosphaeridia atava Naumova 1960 [73], Plate 3, figs. 15; Yankauskas *et al.*, 1989 [61], pp. 74–75, Plate X, figs. 4–7

Holotype: paleontological collection of the Institute of Precambrian Geology and Geochronology, St. Petersburg, Russia, slide no. 452/1; Upper Riphean, Lakhanda Group, Neryuen Formation, Uchur–Maya region, Siberia (lost).

Description. Solitary spheroids with one-layer translucent envelope; outer surface and walls of spheroids are fine-grained. The walls are about 1.5 μm thick. Spheroids are about 130 μm in diameter.

Comparison. Spheroids *L. atava* have different diameter and wall thickness as compared to other *Leiosphaeridia* species (see [61], pp. 24–25, Plate 3).

Distribution: is common in many assemblages of Proterozoic microfossils.

Material: one specimen.

Leiosphaeridia crassa (Naumova) emend. Jankauskas, 1989

Plate III, figs. 4–5

Leiosphaeridia crassa Naumova, 1949 [73], Plate I, fig. 3; Yankauskas *et al.*, 1989 [61], pp. 75-76, Plate IX, figs. 5–10.

Holotype: Naumova, 1949, Plate I, figs. 3; paleontological collection of the Institute of Precambrian Geology and Geochronology, St. Petersburg, Russia, preparation no. 452/1; Lower Cambrian, Lontova Formation, Estonia.

Description. Solitary spheroids with one-layer opaque envelope; their outer surface and walls are coarse-grained or granular. The walls are more than 2.0 μm thick. Spheroids range in diameter from 27.0 to 68.0 μm.

Comparison. Spheroids *L. crassa* have different diameter and wall thickness as compared to other *Leiosphaeridia* species (see [61], pp. 24–25, Plate 3).

Distribution: the species is a component of many assemblages of Proterozoic microfossils.

Material: more than 50 specimens.

Genus *Pterospermopsimorpha* Timofeev, emend. Mikhailova et Jankauskas, 1989

Type species: *P. pileiformis* Timofeev (Timofeev, 1966).

Pterospermopsimorpha pileiformis Timofeev, emend. Mikhailova, 1989

Plate III, figs. 1–3, 9

Pterospermopsimorpha pileiformis [74]; Yankauskas *et al.*, 1989 [61], pp. 49–50, Plate III, figs. 7, 8.

Holotype: paleontological collection of the Institute of Precambrian Geology and Geochronology, St. Petersburg, Russia, slide no. 453/2; Upper Riphean, Lakhanda Group, Neryuen Formation; Uchur–Maya region, Siberia (lost).

Lectotype: paleontological collection of the Institute of Precambrian Geology and Geochronology, St. Petersburg, Russia, slide no. 49/21-X1-72; Upper Riphean, Miroedikha Formation; Turukhansk Uplift, Siberia.

Description. The outer envelope is compact, slightly wrinkled, with a shagreen surface; it is 45 to 100 μm in diameter. The more compact internal body with shagreen surface is thick, slightly translucent, 30–60 μm in diameter, having a smooth distinct contour.

Comparison. In distinction from other species of the genus, the studied specimens have shagreen outer envelopes wrinkled a little and smoothly sculptured.

Distribution: Riphean and Vendian deposits of Northern Eurasia.

Material: ten variably preserved specimens.

Genus *Granomarginata* Naumova, 1960

Granomarginata ? sp.

Plate III, figs. 8, 10, 11

Description. Solitary spheroids with two-layer envelope, the inner layer of which is surrounded, as if it is rimmed, by the less dense layer; walls of the inner spherical translucent or opaque layer are about 1.5 μm thick and sometimes fractured. The outer translucent layer is also of spherical shape, fine-grained, 1.0 to 1.5 μm thick. External and internal diameters of spheroids range from 35.0 to 40.0 μm and from 30.0 to 40.0 μm, respectively.

Remarks. Some microfossils described as *Granomarginata ?* sp. may characterize the other, partially decomposed taxa, for instance *Pterospermopsimorpha pileiformis*, while other specimens are slightly altered remains of planktonic microorganisms, which are similar to some two-layer morphotypes from younger deposits.

Material: several variable preserved specimens.

Unnamed Planktonic Forms

Plate III, fig. 7

Description. A solitary spheroid with multilayer envelope, the inner translucent layer of which is rigid, more or less spherical in shape; walls of this layer are about 1.5 μm thick. Outer transparent layers are of regular wavy outlines. Their external surface is fine-grained, and walls are less than 1.0 μm thick. External

and internal diameters of spheroids are approximately as large as 55 and 45 μm, respectively. Between the outer and inner envelope layers, there are seen problematic processes, the light tubular structures about 5.0 μm long, which range in diameter from 2.0 to 3.0 μm.

Remarks. The studied specimen has something like spine at the surface (Plate III, fig. 7). The material is not representative, however, inappropriate for concrete biological inferences and taxonomic classification. The specimen may represent either a phytoplankton form having a primary process, or an altered cyanobacterial remain of the *Gloeodiniopsis* type.

Material: one specimen.

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Correlation of Frasnian–Famennian Boundary Deposits in Western Regions of the Magnitogorsk Megazone (Eastern Flank of Southern Urals)

G. A. Mizens, V. V. Chernykh, and L. I. Mizens

Institute of Geology and Geochemistry, Ural Division, Russian Academy of Sciences, Pochtovyi per. 7, Yekaterinburg, 620151 Russia

Received December 26, 2002

Abstract—Upper Devonian clastic complexes (the Zilair Formation flysch, Biyagoda mixtite, Koltuban limestone conglomerate, Bugodak volcanogenic–sedimentary sequence) are considered. In the eastern flank of the Southern Urals, the Zilair Formation is shown to be represented by two sequences similar in appearance but different in composition, consisting of clastic material derived from different sources. Accordingly, it is reasonable to rise the subdivision rank up to a group (by the analogy with that in the western slope of the Urals) combining two flysch formations (which are referred to as "Prisakmara" and "Bol'shoi Kizil") and underlying mixtites and conglomerates. The Karantau flysch underlying the Biyagoda mixtite represents a part of the Sakmara Formation that formed due to erosion of the Uraltau cordillera. Based on conodont and brachiopod faunas, lumpy limestone conglomerates (the Koltuban sequence) are proved to constitute only one level of rudaceous rocks. Complicated interrelations between calcirudites and cherts of the Mukasovo Formation are accounted for the slumping origin of conglomerates. It is shown that limestone fragments in these conglomerates range in age from the middle Frasnian to earliest Famennian. The conglomerates formed during the period corresponding to the second half of the conodont *triangularis* Zone and initial *crepida* Zone, when the global sea-level drop caused destruction of shelf limestones in an island-arc inactive area. The formation of the Biyagoda mixtite was related to the same event, though under conditions of active volcanism in the island arc. The Biyagoda mixtite and silicate components of the Koltuban conglomerates are fully analogous in composition to the Bol'shoi Kizil Formation overlying them. These subdivisions are genetically interrelated, and the unnamed rudaceous sequences also belong therefore to the Zilair Group. The lower boundary of the group corresponds to the top of the Mukasovo Formation (the conodont *rhenana* Zone). Accumulation of the Sakmara Formation commenced prior to deposition of the Bol'shoi Kizil Formation in the middle of the *triangularis* hemera, i.e., at the incipient moment of the *linguiformis* time.

Key words: Frasnian, Famennian, conodonts, brachiopods, Southern Urals, Mukasovo Formation, Zilair Formation, mixtites, conglomerates, cherts, regression, eustatic changes.

INTRODUCTION

Upper Devonian sedimentary complexes occupy vast areas in the western and eastern flanks of the Southern Urals (Fig. 1), being over 3000 m thick. They are therefore of exclusive importance for reconstruction of the regional geological history. Nevertheless, stratigraphic problems to be resolved with regard to these complexes are numerous. The situation reflects the clastic nature of mentioned rocks, which naturally yield poor organic remains. In such a case, lithostratigraphy is of great significance along with biostratigraphic methods.

Much new data on stratigraphy and lithology of the Zilair Formation and underlying beds was obtained (e.g., [1]) and nevertheless, the entire range and boundaries of Upper Devonian complexes, including those of the Magnitogorsk megazone discussed in this work, are still not fully understood.[1] The situation is more or less satisfactory at present with the Mukasovo Horizon (Formation, Sequence). Investigations of V.A. Maslov and his colleagues [2] in western regions of the Magnitogorsk megazone (the Western Magnitogorsk and Voznesensk–Sakmara zones) showed that cherts and mudstones of the Mukasovo Formation correspond in age to the *punctata, hassi, jamieae, rhenana* conodont zones. The zones span the Domanik and Mendym horizons and a part of the Askyn Horizon of the Frasnian, the composition and thickness of which considerably change in lateral direction. On the north (north of the

[1] With respect to deposits in the eastern flank of the Southern Urals, the term "Zilair Formation" is generally applied. The term "group" is in use with respect to analogous deposits of the western flank (Zilair synclinorium), where several formations are distinguished, or with respect to Upper Devonian sequences of the Southern Urals in general.

Khudolaz River latitudinal course), the Mukasovo Formation 600 to 800 m thick includes interlayers and members of volcanogenic–sedimentary rocks, whereas in the south it is condensed, almost deprived of volcanogenic material, and its thickness usually does not exceed several tens of meters. The situation with origin and correlation of thick rudaceous complexes separating the Mukasovo cherts from of the Zilair flysch in the Western Magnitogorsk zone is intricate so far, and, hence, the position of the Zilair Formation lower boundary remains uncertain in the eastern flank of the Southern Urals. In different areas, this boundary is placed at different levels [3] that is unsuitable for correct correlation and leads to confusing views on the terrigenous sedimentation development and on the appearance time of particular provenances.

Within the Voznesensk–Sakmara zone, like in the Gai region of the Western Magnitogorsk zone, the Zilair Formation is usually assumed to rest directly on the Mukasovo siliceous shales. The Zilair and Mukasovo formations are separated from each other by diverse deposits, e.g., by rudaceous limestone conglomerates (table) in the Baimak–Buribai and Sibai areas or by the sandy–clayey Karantau Formation and Biyagoda mixtite in the basin of Bol'shoi Kizil River. Volcanic rocks of the Novovoronino Formation and volcanogenic mixtite of the Bugodak sequence separate them near the Town of Verkhneural'sk and along the Uzunkyr Range, while volcanogenic mixtite and, in places, laminated deposits of the Bugodak sequence occur between the mentioned units in the Uchaly region. Stratigraphic position of the Zilair Formation lower boundary irregularly varies from the base of the *linguiformis* conodont zone (where it is assumed to rest directly on the Mukasovo cherts) to the upper subzone of the *crepida* Zone. Clastic complexes occurring between the Mukasovo cherts and the Zilair Formation are ambiguously interpreted by researchers. The nature of some of them is still ambiguous, the age and boundaries, and even the thickness ranges are often unclear.

GENERAL CHARACTERISTICS OF CLASTIC FRASNIAN–FAMENNIAN BOUNDARY SEQUENCES IN WESTERN AREAS OF THE MAGNITOGORSK MEGAZONE

In southern regions, where the Mukasovo Formation is of a small thickness, the so-called Koltuban (mostly reefal) limestones are widespread. These rocks described by L.S. Librovich as early as in the 1930s are considered even in recent works as bodies occurring *in situ* and associated in places with lenses of limestone conglomerate, although their clastic origin has been suggested long ago [5]. The allogenic origin of the Koltuban limestones has been confirmed by facies analysis [6]. Actually, these are boulder–pebble conglomerates with blocks and olistoliths (with longer axes extending for tens to hundreds meters) of massive reefal and laminated limestones. As is shown below, we also found

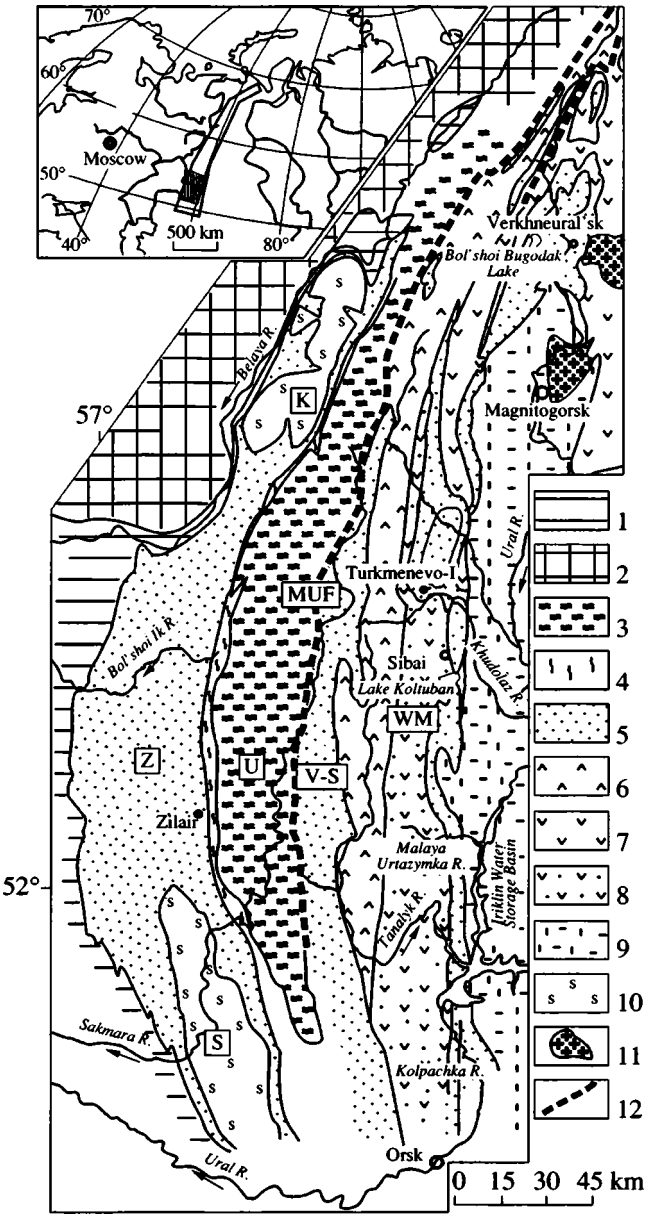
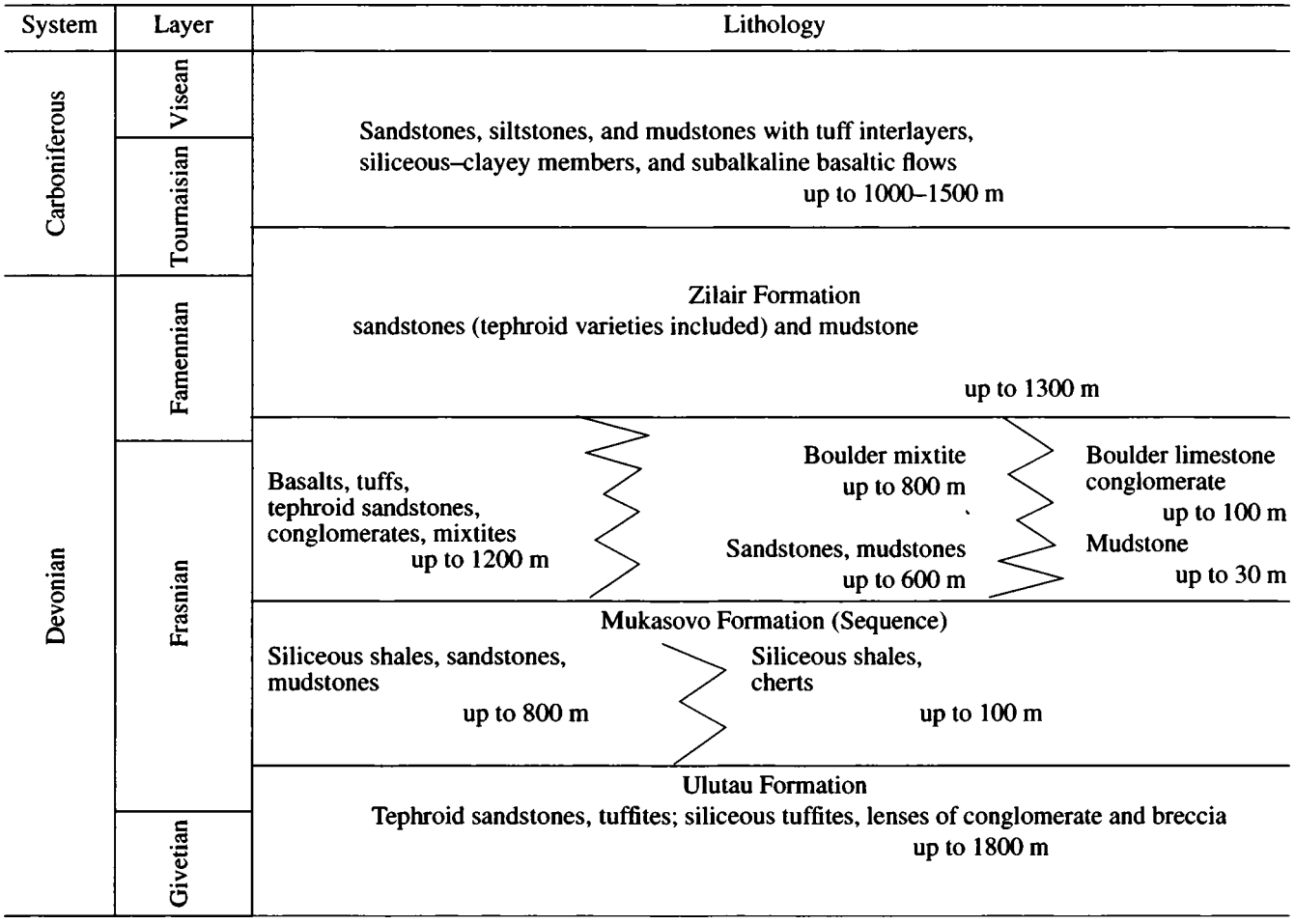


Fig. 1. Geological sketch map of the western segment of the Southern Urals (based on the Geological Map of the Urals, scale 1 : 500000, 1979): (1) Middle–Upper Paleozoic deposits of the Russian Platform margin and Uralian Foredeep; (2) Riphean deposits of the Bashkirian meganticlinorium; (3) Lower–Middle Paleozoic metamorphic complexes of the Uraltau antiform; (4) Silurian–Devonian siliceous sequence; (5) Upper Devonian clastic rocks of the Zilair Group; (6) Eifelian volcanogenic complexes (the Irendyk island arc); (7) Givetian and Upper Devonian volcanogenic complexes, the Magnitogorsk island arc formation included; (8) Middle–Upper Devonian volcanogenic–sedimentary rocks (the Ulutau Formation); (9) Carboniferous sedimentary and volcanogenic deposits; (10) allochthons of the Sakmara zone and Kraka massifs; (11) intrusive granitoid massifs; (12) ultramafic melange zones; (Z) Zilair synform; (U) Uraltau zone; (V-S) Voznesensk–Sakmara and (WM) Western Magnitogorsk zones of the Magnitogorsk megazone; (S) Sakmara zone; (K) Kraka massifs; (MUF) Main Urals Fault (the study area is hatched in the inset map).

Lithostratigraphy of Upper Devonian–lowermost Carboniferous deposits in the Western Magnitogorsk zone of the eastern flank of the Southern Urals (modified after Artyushkova and Maslov [4])



paleontological evidence for clastic origin of limestones under consideration. Indications of this kind were known previously and left without due attention. In addition to limestone pebbles and boulders, conglomerates enclose a variable amount of fragmented volcanic rocks and underlying cherts, as well as few interlayers of limy (more rarely polymictic) turbidites.

Conglomerate beds seem to be not continuous, and their thickness varies from 0 to 100 meters. They are mainly rest on the Mukasovo cherts or locally replace the latter along the strike. Only in the southernmost area (the Gai and Orsk regions), the clayey–sandy strata up to 400 m thick overlie the cherts, which are overlain in turn by polymictic conglomerates with limestone fragments and blocks, which yield the Askyn conodonts and brachiopods, including those of the *triangularis* Zone ([7]. The described conglomerates seem to be related in origin to destruction of shelf limestone of the island-arc chain on the east [8].

North of the Khudolaz River latitudinal course, in the Bol'shoi Kizil River basin, cherts of the Mukasovo Formation are overlain by sandy–clayey deposits of the Karantau after Nestoyanova [9] or Kodinka after Smirnov [10] Formation that is up to 500–600 m thick [4]. Its thickness decreases down to 10–20 m eastward (in sections of the eastern limb of the Khudolaz syncline). Deposits of the Karantau Formation are similar in appearance to the classic Zilair sediments occurring westward and above the Biyagoda olistostrome higher in the section. That is why some researchers [11] suggested the tectonically doubled section of the Zilair Formation in that region owing to thrusting at the olistostrome base. As is established later however, the rock succession is undisturbed here. The Karantau Formation yields conodonts of the *linguiformis* Zone, while sandstone and mudstone above mixtites bear conodonts of the upper subzone of the *triangularis* Zone [4]. The contact between deposits of the Karantau Formation and mixtites is normal sedimentary [8]. The studied sandstone composition (Fig. 2) suggests that the Karantau Formation is composed of clastic components of western origin characteristic of the Zilair deposits (fragments of metamorphic rocks, various volcanites, mica, serpentinite, etc.).

The Biyagoda mixtite, like the Karantau Formation directly underlying it, is exposed in the Bol'shoi Kizil

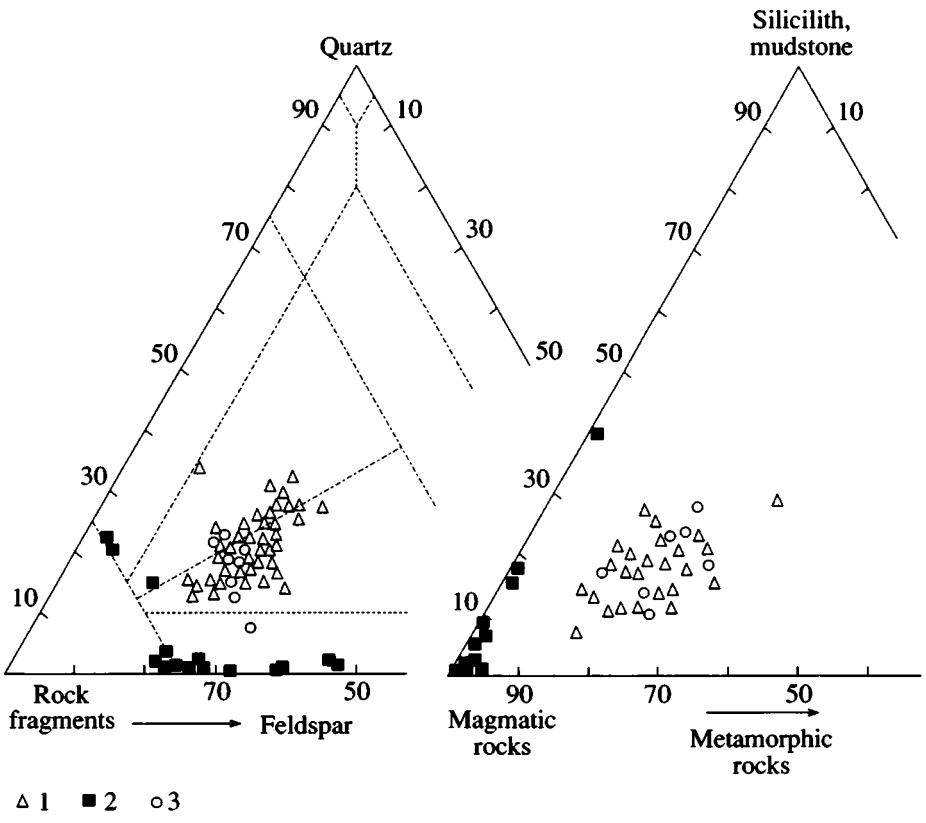


Fig. 2. Composition of upper Frasnian and lower Famennian sandstones in western regions of the Magnitogorsk megazone: (1, 2) sandstones of the Zilair Formation genetically related to (1) the Uraltau cordillera (Ziren-Agach, Bol'shoi Buzavlyk, Khudolaz, Malaya Urtazymka, Iriklin water storage basin, Ishmurzino, Nigamatovo, and Bol'shoi Bugodak sites) and (2) island arc (Kolpachka, Bol'shoi Kazmash, Ikstimer, Malaya Urtazymka, Uzel'ga, Turkmenevo-I, and Lake Koltuban sites); (3) sandstones of the Karantau Formation (Biyagoda Range, Turkmenevo-I, and Ikstimer River sites).

River basin only. Mixtite has been traced along the strike for about 50 km as a fill of the so-called Biyagoda (or Khudolaz) syncline. Mixtite, termed by Smirnov *et al.* [10] as the Biyagoda olistostrome, represents an unsorted sandy–clayey mass with irregularly distributed boulders and fragments of volcanic rocks (mainly basic), cherts, sandstone, and limestone. Individual blocks are over 1000 m in size. The thickness of described mass deserves a special notice. For some preposterous mistake, first researchers underestimated the thickness (sometimes by the order of magnitude), and the mistaken assessments traveled later from one work into another. According to our calculations, the true thickness of mixtites is up to 700–800 m in places. The mixtite unit was likely formed in response to the collapse of tephra and lava deposits in the coastal zone of volcanic islands [8].

At the latitude of Verkhneural'sk and northward to the Uchaly, the Bugodak volcanogenic–sedimentary sequence up to 1200 m thick overlies cherts of the Mukasovo Formation [12]. Basic volcanogenic rocks (lava included) dominate in the sequence, while mixtite with the volcanic, silicilith and limestones blocks are present, like rare sandy–clayey members, as intercalations. Laminated rocks likely dominate in western sections. On the Ural River right bank at the latitude of Ural'sk (south of the Uchaly), the sequence 300–400 m thick is composed of alternating polymictic sandstone, mudstone and gravelstone considerably increasing in thickness northeastward.

In the eastern flank of the Urals, like in the western one, the Zilair Formation is represented by alternating sandstone and mudstone beds. Most researchers class these deposits with flysch. The Zilair Formation of the Magnitogorsk megazone is clearly divisible in two different sequences [13], replacing each other via facies alternation in lateral direction. One sequence up to 1500 m thick is composed of diverse clastic material (Fig. 2) represented by fragments of acid to basic volcanics, metamorphic rocks, serpentinites, cherts, and by mineral grains of quartz, plagioclase, epidote, mica (muscovite, stilpnomelane, biotite), Cr-spinel, garnet, zircon, and various Ti-bearing minerals. This sequence is persistent in composition along the strike and vertically, being similar to lithologic counterparts in the western flank of the Urals (Zilair synclinorium). The sequence is likely related in origin to the cordillera erosion in the area of the present-day Uraltau massif. Clas-

tic material of the other sequence up to 500–600 m thick was most likely derived from an active island arc on the east, like that of mixtite and conglomerate underlying it. The mineral composition of sandstones is less diverse: they are composed mainly of volcanic rock fragments. In places, grains of sedimentary rocks (limestones, cherts) play an important part. Mineral grains are those of plagioclase, quartz, clinopyroxene, rare hornblende and biotite. This sequence is laterally not persistent in composition. Fragments of basic volcanics dominate in some areas, whereas clasts of acidic rocks prevail in other regions. Despite many sections exhibiting alternation of beds, which are genetically connected with different provenances, the material derived from the Uraltau cordillera is characteristic of western regions, while the material of volcanic islands is common in eastern areas.

The basin, in which deposits of the Zilair Formation and underlying rocks accumulated, was of the back-arc type [13]. In the Middle and initial Late Devonian (in the Frasnian), an active island arc was a main source of volcaniclastic and pyroclastic material that accumulated in the Ulutau volcanogenic–sedimentary formation and in of sandstone (volcaniclastic, tephroid) and tuff members of the Mukasovo Formation. Two provenances (an island arc and the Uraltau Cordillera) presumably existed since the end of the Frasnian. Nevertheless, some distinctions in composition and structure of sedimentary and volcanogenic–sedimentary complexes, which are recorded beginning from the Mukasovo level, suggest that sedimentary basin was separated in two provinces of sedimentation (southern and northern) by the boundary now traceable along the Khudolaz River latitudinal course.

RELATIONS BETWEEN BOULDER CONGLOMERATES AND CHERTS OF THE MUKASOVO FORMATION

As mentioned above, boulder limestone conglomerates up to 100 m thick, which overlie or laterally replace cherts of the Mukasovo Formation, are characteristic of Upper Devonian sections in the southern province. Limestones of the Askyn age, which are younger than deposits of the Mukasovo Formation, are dominant rock types in boulders and pebbles (extended olistoliths included) constituting conglomerates [14]. However, the lateral replacement of cherts by conglomerates and limestone conglomerate lenses found recently in the cherts [15] imply uncertainty with respect to number of conglomerate levels. It seems doubtful that there is only one Askyn sequence of clastic limestones, which overlie cherts of the Mukasovo Formation.

To get a deeper insight into the problem, we studied facies relationships between rudaceous rocks and cherts, and determined ages of limestone blocks in sections of different structures. We focused attention on the basin of Malaya Urtazymka River (a right tributary of Ural River), where cherts and limestone conglomerates are exposed in all possible combinations (Fig. 3). In the river valley, one may observe conglomerates directly overlying deposits of the Ulutau Formation and bearing in places the fragments and large detached masses of Mukasovo cherts, which are folded sometimes, as well as volcanogenic–sedimentary rocks with pebbles and boulders (Site 2675). In the other exposure (Site 2674), a block of laminated bleached cherts crowns the conglomerate section. In the southern watershed, conglomerates and associated boulders rest directly above cherts (Site 2676). Neither cherts, nor conglomerates were found in the section of the valley left wall, west of the Alekseevka Village (Site 2673), where Zilair sandstones overlie deposits of the Ulutau Formation.

Contacts between cherts and conglomerates (or between cherts and large limestone bodies) are always abrupt, without a sign of gradual transition. Cherts were likely deposited in deep-water environments. These are fine- to micro-laminated rocks exposed in a vast territory. They yield radiolarians, conodonts, and fine plant detritus. Limestones (pebbles, boulders, and lumps) contain typical shallow-water macro- and microfauna. In addition to carbonate clastic material, conglomerates bear pebbles and grains of silicate rocks, mainly volcanogenic. Interlayers of carbonate and polymictic turbidites are present in places. All these features imply the allogenic origin of limestones: the boulder conglomerates were deposited from debris flows in a deep-water basin.

Conodont assemblages found in limestone boulders are as follows: (1) *Ancyrodella buckeyensis* Stauffer, *A. nodosa* Ulrich et Bassler, *Polygnathus brevis* Miller et Youngquist (*jamieae–rhenana* zones, uppermost part of the Domanik–Mendym Horizon), Site 2675 above cherts; (2) *Palmatolepis winchelli* (Stauffer), *Pa. foliacea* Youngquist, *Ozarkodina nonaginta* Klapper, Kuz., et Ovnat. (*rhenana* Zone, the Mendym Horizon), Site 2675 above cherts; (3) *Palmatolepis* cf. *triangularis* Sannemann, *Pa. quadrantinodosalobata* Sannemann (lower subzone of the *crepida* Zone), Site 2674, below cherts.

Brachiopods were also collected from several boulders. One their assemblage (Site 2674 below cherts) consists of *Schizophoria* sp., *Praewaagenoconcha* sp., *Spinulicosta* sp., *Hypothyridina* cf. *crassicostata* Nal., *Parapugnax* sp., *Pugnoides triaequalis* Mark., *Eoparaphorhynchus lentiformis* (Nal.), *Phlogoiderhynchus* ex gr. *formosa* (Schnur), *Spinatrypina?* sp., *Gibberosatrypa gibberosa* (Mark.), *Adolfia zickzack* (Roem.), *Undispirifer* sp., *Cyrtospirifer tschernyshewi* Khalf., *Cyrtospirifer* sp., *Athyris angelica* Hall, *Ath. bayeti* Rig., *Ath. concentrica* Buch, and *Ath.* ex gr. *globularis* Phill. Fossils are characteristic of the Barma Beds in the western flank of the Urals, the age of which corresponds to the uppermost Frasnian–lowermost Famennian [16]. The other assemblage (Site 2675 below

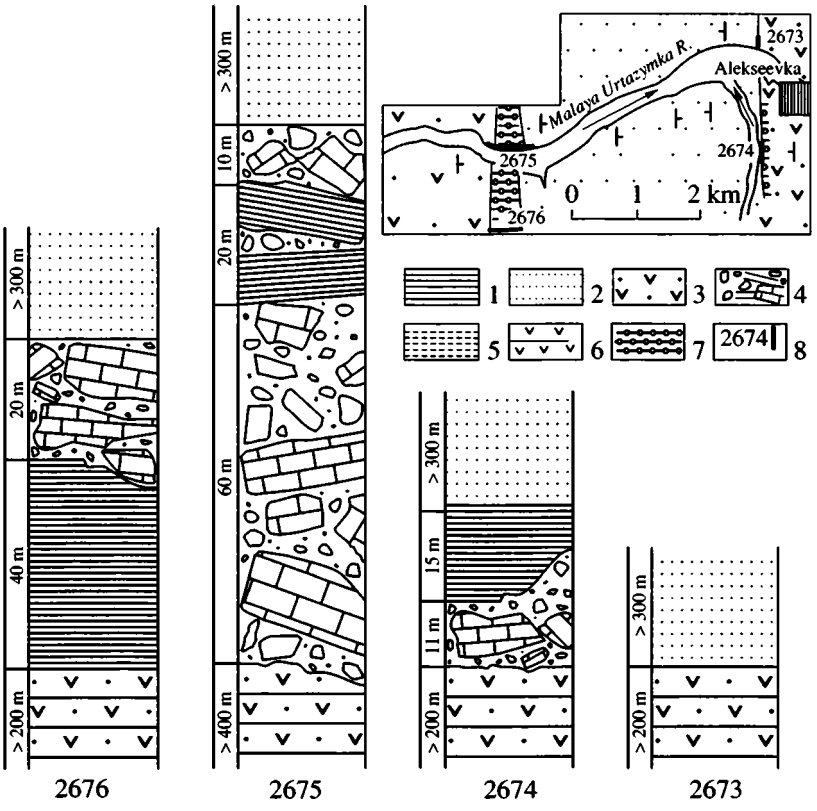


Fig. 3. Relationships between boulder limestone conglomerates and cherts of the Mukasovo Formation in sections along the Malaya Urtazymka River; site localities are shown in the inset map: (1) cherts and mudstones of the Mukasovo Formation; (2) sandy–clayey sediments; (3) volcanogenic–sedimentary deposits of the Ulutau Formation; (4) mixtites and conglomerates; (5) mudstones with siltstone interlayers; (6) volcanics (see Fig 4); (7) cherts of the Mukasovo Formation and limestone conglomerates (undifferentiated); (8) sites numbers.

cherts) is represented by *Gypidula* cf. *biplicatiformis* Mark., *G.* ex gr. *brevirostris* (Phill.), *Devonogypa* ex gr. *globa* (Bronn), *Hypothyridina* cf. *incisiva* (Roem.), *Spinatrypina bifurcata* (Mark.), *Iowatrypa* sp., *Carinatina? biohermica* Yud., *Adolfia* ex gr. *zickzack* (Roem.), *Warrenella* cf. *koltubanica* (Nal.), *Emanuella* cf. *subumbona* (Hall), and *Cyrtospirifer* sp. (the Askyn Horizon of the upper Frasnian Substage). The third assemblage (Site 2675 above cherts) includes *Aulacella?* sp., *Schizophoria* sp., *Gypidula* ex gr. *brevirostris* (Phill.), *Productella* sp., *Hypothyridina* cf. *cuboides* (Sow.), *Pseudoatrypa posturalica* (Mark.), *P.* aff. *posturalica* (Mark.), *Adolfia zickzack* (Roem.), *Emanuella subumbona* (Hall), *Conispirifer* cf. *conoideus* (Roem.), *Cyrtospirifer* sp., and *Pyramidalia simplex* (Phill.) (the Askyn Horizon of the upper Frasnian Substage). Finally, brachiopods from Site 2675 above cherts are *Aulacella* sp., *Gypidula* ex gr. *biplicata* (Schnur), *Devonogypa* ex gr. *globa* (Bronn), *Hypothyridina crassicostata* Nal., *Pseudoatrypa* cf. *posturalica* (Mark.), *Desquamatia* sp., *Iowatrypa* ex gr. *kadzielniae* (Gür.), and *Cyrtospirifer* sp. (the Askyn Horizon of the upper Frasnian Substage).

Thus, regardless of the conglomerate position above, below, or within the Mukasovo cherts in the studied area, the age of limestones boulders corresponds to the stratigraphic interval from the upper part of the Domanik Horizon to the lower part of the Makarovo Horizon inclusive (the *jamieae* conodont zone–lowermost *crepida* conodont zone). In other words, conglomerates contain fragments of limestones derived from the Mukasovo level, the range of which corresponds to interval of the Domanik–Askyn horizons [4], and younger rocks, those of the lowermost Famennian included.

Data on sections in other areas confirm the above age determinations. The middle to upper Frasnian and basal Famennian faunas (the *rhenana, linguiformis,* and *triangularis* conodont zones) were detected in limestone bodies near the Lake Koltuban [17]. In the section Mukasovo-1, the age of limestone boulders corresponds to the *linguiformis* Zone [18]. Polymictic conglomerates in the region of the Tanalyk River lower course (Tashtugai Creek) contain abundant limestone boulders with Askyn brachiopods [1]. Limestone interlayers (?) in the conglomerate sequence of the Kolpachka River mouth yield conodonts and brachiopods

of the *triangularis* Zone [19]. Smirnov *et al.* [10] reported numerous brachiopod taxa of the Askyn age from limestone boulders widespread in the last sequence.

In many sections (especially where conglomerates rest directly on volcanogenic–sedimentary deposits of the Ulutau Formation), conglomerates enclose cherts as fragments, blocks, and packets of intensely folded beds. This can be observed in the Malaya Urtazymka River basin, in exposures along Tanalyk River (Tashtugai and Malyatino sites) and in other areas as well. For instance, Maslov *et al.* [15] pointed out that limestone conglomerates of the Mukasovo-1 section contain a great amount of fragmented cherts of the Mukasovo Formation, and above this level there is a member of intensely folded cherts with conodonts of the upper part of the Mukasovo Horizon (*Ancyrodella* sp., *Palmatolepis* cf. *jamieae* Ziegl. et Sandb., *Pa.* cf. *rhenana nasuta* Müll., *Polygnathus* aff. *decorosus* Stauffer).

As already noted, the conglomerates under discussion are, as a rule, not monomictic. They contain, along with silicilith fragments, a various amount of pebbles and grains of volcanic rocks (more often acid), as well as grains of quartz and plagioclase. In places (the Kolpachka River mouth), silicate material prevails over carbonates. Sometimes, for instance, in the section along the Tashtugai Creek, conglomerates composed of limestone pebbles and boulders contain interlayers of polymictic sandstone and gravelstone resembling turbidites. Lower beds of the Zilair Formation, which overlie in the study region the limestone and affiliated polymictic conglomerates, are composed of siliciclastic material similar to that of conglomerates. Beds of clastic limestones are developed at the base of this formation in many sections (especially in area of the Iriklin water storage basin) that is indicative of a gradual transition between the limestone conglomerates and graywackes of the Zilair Formation.

The age of lower beds of the Zilair Formation was established in three sites (2673, 2674, 2675) along the Malaya Urtazymka River. Conodont assemblages at all the sites correspond to the upper part of the *crepida* Zone. The identified taxa are (1) *Palmatolepis minuta minuta* Branson et Mehl, *Pa. minuta loba* Helms, *Pa. crepida* Sannemann, *Pa. quadrantinodosalobata* Sannemann; (2) *Palmatolepis minuta loba* Helms, *Pa.* cf. *crepida* Sannemann; and (3) *Palmatolepis tenuipunctata* Sannemann, *Pa. quadrantinodosalobata* Sannemann, *Icriodus alternatus alternatus* Branson et Mehl, *Polygnathus praecursor* Matyja. The *crepida* Zone is likely to exist also in southern sections of the Iriklin water storage basin, on its left bank in the area of the Solenaya River mouth, where we found *Palmatolepis glabra elongata* Holmes, *Pa. glabra pectinata* Ziegler, *Pa. glabra glabra* Ulrich et Bassler, *Pa. superlobata superlobata* Branson et Mehl, *Pa. superlobata helmsi* Ovnatanova, and *Pa. quadrantinodosalobata* Sannemann. Stepanova *et al.* [19] confirm this age interpretation. Lower horizons of the Zilair Formation yield conodonts of the *crepida* Zone on the right bank of the Iriklin water storage basin as well, in sections above and below the Tanalyk River mouth [20]. The upper *triangularis* Subzone is established at similar level along Tashtugai, Tulubai, and Kolpachka rivers [21].

Brachiopods collected in the area of the Solenaya River mouth are also indicative of the lowermost Famennian. Their assemblages are as follows: (1) *Schizophoria* ex gr. *impressa* Hall, *Praewaagenoconcha* sp., *Camarotoechia* cf. *turanica* (Rom.), *Cyrtospirifer* sp., *C.* cf. *sulcifer* (Hall et Clarke), *Mucrospirifer* sp., *Athyris concentrica* Buch.; (2) *Schizophoria* sp., *Praewaagenoconcha* ex gr. *speciosa* (Hall), *Parapugnax* sp., *Camarotoechia turanica* (Rom.), *Mucrospirifer* sp., *M.* ex gr. *posterus* (Hall et Clarke), *Spinospirifer* sp., *Cyrtospirifer* cf. *calcaratus* (Sow.), *C.* cf. *lebedianicus* (Nal.), *C.* cf. *trapezoidalis* Krest. et Karp., *Athyris concentrica* Buch.; (3) *Schizophoria* sp., *Praewaagenoconcha* ex gr. *speciosa* (Hall), *Parapugnax* sp., *Cyrtospirifer* aff. *brodi* Wenjuk., *C.* cf. *trapezoidalis* Krest. et Karp., and *C.* cf. *calcaratus* (Sow.)

AGE OF CONGLOMERATES AND MIXITES

Based on data of the previous section, we may state that boulder limestone conglomerates were deposited within the interval bounded by the conodont *rhenana* Zone (the top of Mukasovo Formation) and the upper subzone of the *triangularis* Zone, in places by the *crepida* Zone (the base of the Zilair Formation), although they also include blocks of older rocks. Is it possible to specify better this interval? In the Gai region, along the Kolpachka Creek (Srednyaya River), cherts of the Mukasovo Formation (Fig. 4) constitute a siliceous–clayey member (more than 10 m thick) with siltstone interlayers that yield conodonts of the *linguiformis* Zone [15]. The overlying sandstone and mudstone with Famennian conodonts are about 400 m thick. This sequence is overlain by conglomerates, limestones boulders of which bear conodonts of the *triangularis* Zone [19]. Conodonts of the *upper triangularis* Subzone are established in sandy–clayey deposits above the conglomerates. Hence, the initial deposition of conglomerates in this area corresponds to the middle of the *triangularis* Zone. A similar situation is characteristic of the Tulubai River site. The "official" Mukasovo Beds are overlain here by fine-laminated mudstones up to 25–30 m thick. Conodonts of the *linguiformis* Zone were also found in interlayers of organoclastic limestones near the base of the beds [15]. Sandy–clayey sequence with conodonts of the *triangularis* Zone rests on mudstones. The section is lacking conglomerates, but they were found a little eastward, along the Malaya and Bol'shaya Urtazymka rivers. The absence of a thin transitional clay member in other sections, which corresponds to the *linguiformis* Zone, can be a consequence of erosion by debris flows that deposited con-

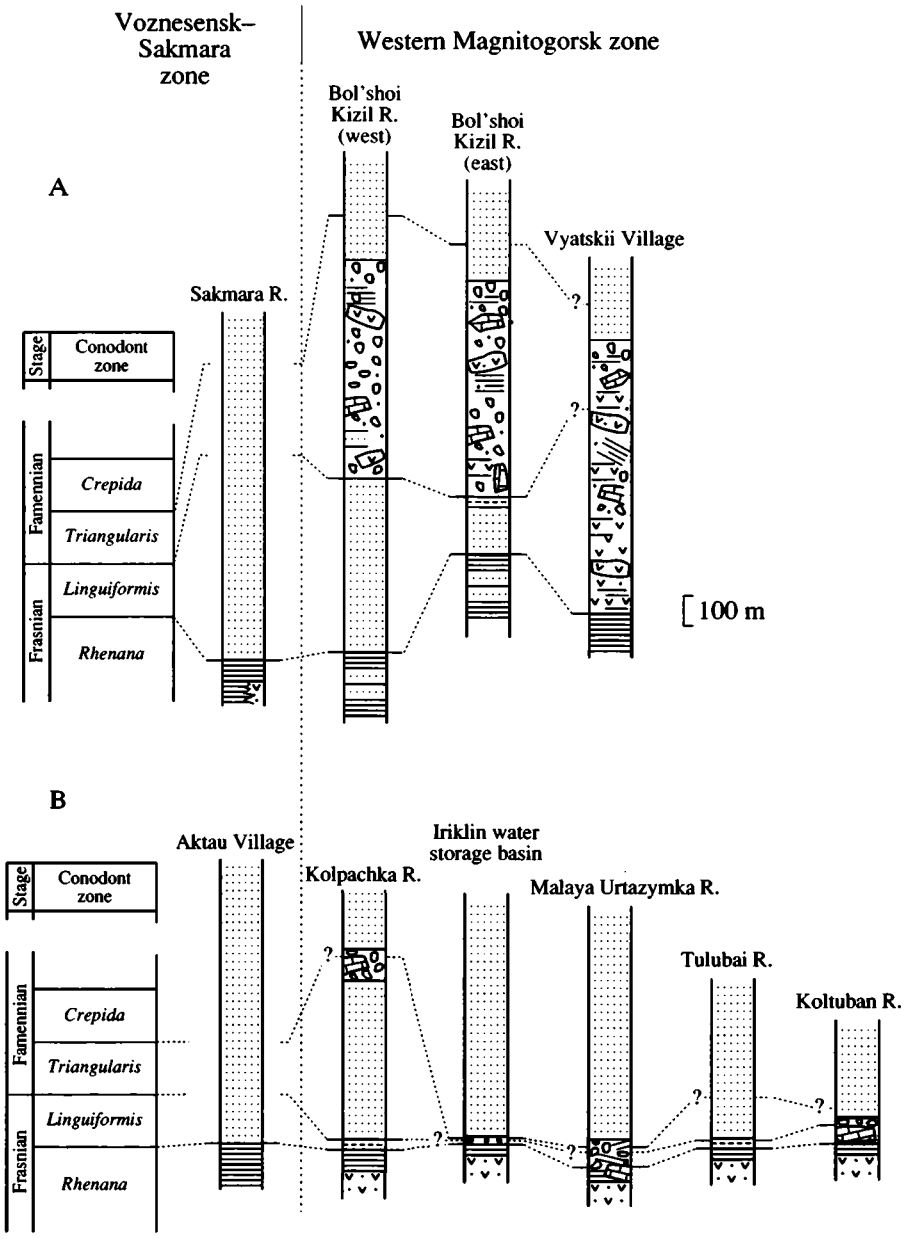


Fig. 4. Correlated clastic sequences above the Mukasovo Formation, the eastern flank of the Southern Urals: (A) northern and (B) southern sections (this study and data of Maslov *et al.* [1]; Artyushkova and Maslov [4]; Maslov *et al.* [15]; symbols as in Fig. 3).

glomerates. Upper, poorly consolidated clayey sediments were destructed. In the first half of the time interval corresponding to the *triangularis* Zone, the turbidite sequence was likely formed in the southernmost part only (Kolpachka River). True enough, it is not clear which part of this zone is present at the base of the turbidite section exposed along the Tulubai River.

The upper boundary of conglomerates (like, probably, the lower one) seems to be diachronous. This is evident from variable ages of lower beds of the Zilair Formation: from the upper subzone of the *triangularis* Zone to the upper subzone of the *crepida* Zone. In addition, limestone boulders ranging in age from lower to middle subzones of the *crepida* Zone are present in some sections. It is possible to suggest that polymictic turbidite flows and flows of clastic material with limestone boulders (or individual boulders) simultaneously run down in different slope canyons over a certain time.

It was not a single act of conglomerate formation. There are individualized interlayers of limestone (Lake Koltuban) and polymictic turbidites (Tashtugai Creek) in the conglomerate complex, although it is hard to speak of particular stages of conglomerate formation. Rudaceous rocks are nowhere separated by background

deposits or relatively thick turbidite packets. Conglomerates do not seem to form a continuous cover. They occur as localized bodies different in thickness and size. If flows of clastic material that is deposited in conglomerates did follow the canyons, then the conglomerate "spots" mark positions of the canyon mouths. Flows, which carried limestone blocks as large as tens and hundreds meters across, completely plowed up the underlying cherts in places, and the mass of carbonate material mixed with fragments of cherty beds. Despite the apparent integrity, large packets of the Mukasovo cherty layers, which are locally observable in sections, are undoubtedly allogenic, if conglomerates underlie or enclose them. For example, in outcrops along the Malaya Urtazymka River (see Fig. 3) or in the section along the Tashtugai Creek it is possible to observe the 5-m-thick member of cherts with conodonts of the *rhenana* Zone, which lies above conglomerates with the Askyn brachiopods [1]. In some cases, a flow of clastic material, the shreds of cherts included, moved down the slope and left behind a barren bottom composed of the Ulutau Formation rocks, which afterward became covered by Zilair deposits, as it can be seen on the left wall of the Malaya Urtazymka River valley (Site 2673) west of the Alekseevka Village.

The Biyagoda mixtite likely formed in the northern province approximately at the same time (the middle–upper part of the *triangularis* Zone; Fig. 4). The mixtite is underlain by sandstones and mudstones of the Karantau Formation, which contain conodonts of the *linguiformis* Zone in the lower part [4], and overlie beds of the Zilair Formation, near the base of which (the Village of Turkmenevo-I) the upper *triangularis* Subzone is recognized [21]. The formation time of the Biyagoda mixtite was undoubtedly short, but the deposition did not represent a single act, because mixtite bodies alternate with individualized turbidite beds like the southerly conglomerates. A possibility that mixtite deposition developed in places, like in the south, during the interval of the lower subzone of the *crepida* Zone cannot be ruled out. The structure and composition of the Biyagoda mixtite show that there were two canyons at least, through which the landslides and debris flows slumped down [8]. The composition of overlying clayey–sandy sequence (the Zilair Formation) is identical to that of the mixtite. Thus, while the sequence formed, the same source of clastic material (volcanogenic complexes of an island arc) was still exposed to erosion.

Analogs of the Biyagoda mixtite are likely present in the Bugodak Sequence (rudaceous tuffs, lavas, tuffites, mixtites) at the latitude of Verkhneural'sk (the section near the Vyatskii Settlement, Fig. 4) and further northward, but the sequence age is poorly substantiated, because no organic remains were found in it. It is possible to suggest only that its lower part corresponds in age to the conodont *linguiformis* Zone, as it directly lies on deposits of the Mukasovo Formation, the top of which corresponds to the *rhenana* Zone [15]. Conodonts of the *Late crepida–rhomboidea* Zone were found in the lower part of the Zilair Formation south of Uchaly [18].

A similar situation is typical of the Berchogur basin in the Mugodzhary area [22]. In the opinion of Veimarn and his colleagues, the olistostrome spans here the interval of the *Late triangularis–crepida* Zone.

Thus, deposition of thick mixtite and conglomerate complexes developed almost concurrently in a vast region from southern and central areas of the Western Magnitogorsk zone (south of Magnitogorsk) to the southern edge of the Urals. The deposition period approximately corresponds to the interval from the middle part of the *triangularis* Zone to the beginning of the *crepida* Zone. As northern regions are poorly studied so far, the mixtite levels cannot be dated further with confidence.

DISCUSSION

If we take the Mukasovo Formation top for the reference level, then a series of eastern clastic sequences appears to correspond to the lower part of the Zilair Formation of the Voznesensk–Sakmara zone (Fig. 4). These are the transitional silty–clayey member and Koltuban limestone conglomerates, the Karantau Formation, the Biyagoda mixtite, and the Bugodak sequence. Actually, the Zilair Formation of the Western Magnitogorsk zone is correlative only with the upper part of the same Zilair Formation of the Voznesensk–Sakmara zone. Moreover, there are considerable distinctions in their composition. It is amply clear at present that on the eastern flank of the Southern Urals there are two laterally contacting sequences attributed to the Zilair Formation, which are of different range and composition. Both sequences were deposited in one and the same basin, under similar conditions, but not quite synchronously and are composed of clastic material derived from different provenances. Within the Voznesensk–Sakmara zone, the Zilair Formation rests directly on the Mukasovo cherts, and its lower boundary corresponds to the base of the conodont *linguiformis* Zone. Eastward, in the Western Magnitogorsk zone, the sandy–clayey sequence of a similar appearance that is assigned to the Zilair Formation overlies the horizon of mixtites and conglomerates. In southern areas, where the *upper triangularis* level is determined in a few places, it usually begins within the upper subzone of the *crepida* zone. The *upper triangularis* level is likely common in northern sections.

Similar complexes occurring below the conglomerate level are not included as a rule into the Zilair Formation (the Kolpachka, Tulubai, Bol'shoi Kizil rivers, and the right bank area of the Ural River south of Uchaly). The Karantau Formation with conodonts of the *linguiformis* Zone (the Bol'shoi Kizil River basin) is of special importance amidst them. The formation likely merges on the west with the Zilair Formation of

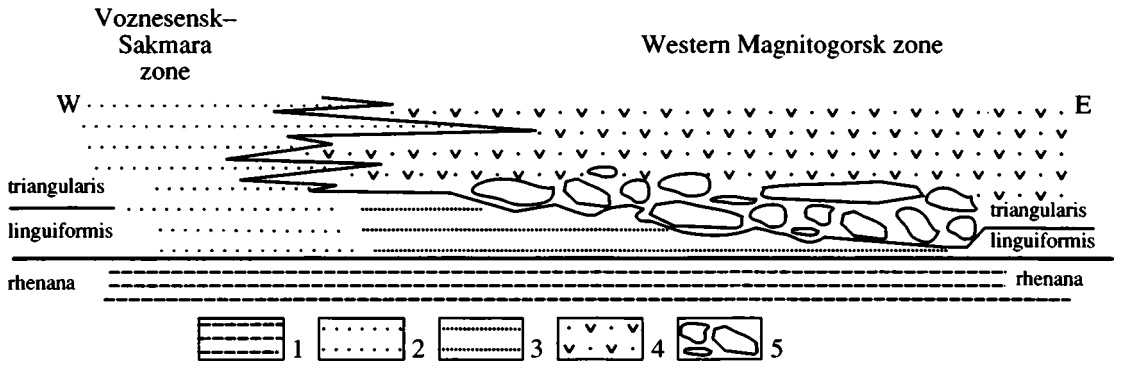


Fig. 5. Sketch illustrating relationships between lower beds of the Zilair and Karantau formations in the Voznesensk–Sakmara and Western Magnitogorsk zones at the latitude of the Bol'shoi Kizil River basin: (1) cherts of the Mukasovo Formation; (2, 4) sandy–clayey deposits of the Zilair Formation genetically related to (2) the Uraltau cordillera and (4) island arc; (3) deposits of the Karantau Formation; (5) Biyagoda mixtite (conodont zones are indicated on both sides of the sketch).

the Voznesensk–Sakmara zone (Fig. 5) as it follows from their similar composition (Fig. 2). The Biyagoda olistostrome occurring above is of a different composition. It is composed of volcanogenic–sedimentary rocks, clastic material of which is derived from an active island arc, whereas the Karantau Formation is formed of material eroded from the Uraltau Ridge. Artyushkova and Maslov [4] include mixtites and the Karantau Formation into one Biyagoda sequence that does not seem reasonable because of different origin of two rock units. On the other hand, the Biyagoda mixtite is closely related in composition and origin to the clayey–sandy sequence overlying it and attributed in this region to the Zilair Formation. The mixtite matrix and sandstones of the mentioned sequence are almost identical in composition. Hence, the Karantau Formation represents eastern facies of that lower part of the Zilair Formation, the origin of which is related to erosion of the Uraltau Ridge. The necessity to attribute the Karantau Formation to the Zilair sequence was pointed out by Smirnov and his colleagues who wrote: "The Karantau Formation is in essence an upper Frasnian part of the Zilair Group section, the further deposition of which was interrupted in this area by a thrust-induced landslide" ([10], p. 132). The Biyagoda mixtite along with deposits overlying represents the Zilair Formation genetically related to destruction of an island arc. The terrigenous influx from the western Uraltau cordillera commenced in the region of the Bol'shoi Kizil River basin and Voznesensk–Sakmara zone soon after the Mukasovo time (the Karantau Formation), whereas clastic material derived from the island arc appeared here in the middle of the *triangularis* Zone.

To the north of Bol'shoi Kizil River, at the latitude of Verkhneural'sk, the Karantau Formation is undistinguishable. Its place in the section is occupied by the lower part of the Bugodak sequence, the upper part of which seems to include analogs of the Biyagoda mixtite. This is only an assumption and nothing else, because organic remains are unknown here. Fossils have not been found as well in overlying deposits of the Zilair Formation, and exact correlation is impossible. The Bugodak Formation is composed completely of material derived from an active island arc, whereas in the Zilair Formation [13] there are intervals of sandstones of the Uraltau origin (Lake Bol'shoi Bugodak) and the eastern volcanogenic genesis (Uzel'ga River). Relationships between these two facies remain unclear so far.

Further to the north, in the Uchaly region, the boundary between the Mukasovo cherts and overlying deposits is rather indistinct, since the Mukasovo Formation includes numerous interlayers of terrigenous rocks (sandstones included). Within the Imangul syncline, this formation is overlain by a sequence similar to that of the Verkhneural'sk region [18] and replaced by laminated rocks in the southward direction. Deposits of the Zilair Formation with Famennian conodonts occur higher in the section. In other words, an intense accumulation of clastic rocks was in progress during the time interval of the conodont *linguiformis* Zone in these (western) sections.

Analogs of the Karantau Formation are also unknown in southern areas of the Western Magnitogorsk zone. In the latest Frasnian, clastic material from the Uraltau did not arrived here. The *linguiformis* Zone is represented in these areas by siliceous mudstone with siltstone interlayers (Kolpachka River) or by shales (Tulubai River). This means preservation of depositional environments of the Mukasovo time.

The synchronism of compositionally different mixtites, outcrops of which mark a belt of considerable extension (over 200 km) in the Magnitogorsk megazone, allows us to think that their deposition was related to some significant event in the basin history. Any link with collision and displacements along major thrust faults, which was surmised formerly [23], is unlikely, as over the whole extension of the belt there is no material in mixtite complexes, which could be considered as foreign to an island arc and its slope. After deposition

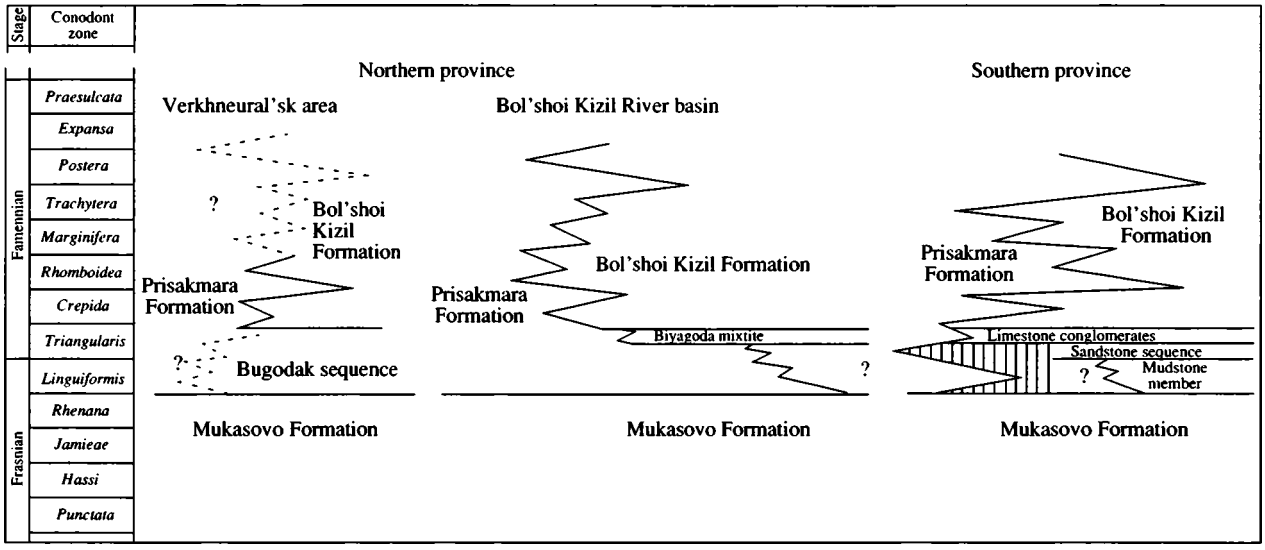


Fig. 6. Correlation scheme for the Zilair Group deposits in various regions of the Southern Urals eastern flank.

of mixtites, sedimentation in the basin did not changed as well. One can assume effect of an intense earthquake, but this is inconsistent with indications of the multiphase slumping, though within a short time period, and with presence of rounded limestone pebbles. On the other hand, it is remarkable that indications of a considerable regression at the Frasnian–Famennian boundary time are recorded in shelf deposits throughout the World. At this time exactly, the relatively deep-sea sediments gave way to shallow-water deposits, conglomerates and breccia appeared, and frequent hiatuses were recorded. These indications are described in numerous works published in Russia [24], Europe [25], Asia [26], North America [27], Australia [28] and used to be related to the sea-level fall, the peak of which was in the middle part of the conodont *triangularis* Zone, right when rudaceous complexes were deposited in the Southern Urals. A large-scale destruction of coastal reefs and the formation of slope breccia and debris-flow deposits have been also characteristic of that time [29]. The described Koltuban conglomerates, which are composed of boulders derived from Frasnian reefs, likely belong to deposits of the last type.

Indications of global regression are well known in the western flank of the Southern Urals as well. The Frasnian–Famennian boundary strata are completely or partially absent in many sections of the platform margin (shelf zone), and deep-sea deposits are replaced by shallow-water sediments in other sections. The regression started here in the *linguiformis* Zone and reached its maximum within the *triangularis* Zone [30]. This is clearly recorded in the carbonate section near the Yaumbaevo Village, where deposits of the lower part of the *triangularis* Zone are absent. In the Bol'shoi Ik River basin, limestones olistoliths in flysch of the Zilair Group also correspond in age to the *triangularis* Zone. A distinct level of thick submarine landslide deposits of the *linguiformis* Zone and almost whole *triangularis* Zone is also distinguished in the western flank of the Central Urals within the basinal sequence of the so-called Kama–Kinel system. The stratigraphy of this sequence was studied in detail by Bikbaev *et al.* [31] in sections along the Chusovaya River.

Eustatic sea-level oscillations across the Frasnian–Famennian boundary were short but rather intense, from 60–100 to more than 150 m according to different assessments [32]. Two conjugate cycles (fall–rise), which lasted 0.1–0.2 Ma each with a break of 0.5 Ma in between, are recognized in places [33]. The first sea-level drop is confined to the uppermost *linguiformis* Zone, the second, most intense fall is recorded in the upper part of the lower subzone of the *triangularis* Zone. Fast and sizable sea-level drops could easily result in destruction of reefs and in slumping of loose deposits accumulated in coastal zones of volcanic islands, which were situated in a deep back-arc basin of the Southern Urals.

After analysis of many published data, Racki [34] concluded that the sea-level fluctuations of Frasnian–Famennian boundary time were controlled by tectonic processes associated with motion of continental plates. His inference seems to be valid. It is likely that there was nothing like intense glaciation at that time. Impact events, similar to that resulted in formation of the grandiose Alamo Breccia of the Late Devonian age in Nevada [35], could influence the sea-level only as a momentary and single act.

CONCLUSION

In the eastern flank of the Southern Urals, the Zilair Formation includes two polymictic clastic sequences different in composition and replacing each other later-

ally. The western sequence is composed of clastic material derived from the Uraltau cordillera, whereas the eastern one is deposited in response to erosion of the volcanic island arc. By analogy with the western flank of the Urals, we propose that the Zilair Formation should be ranked now as the group including two, the Prisakmara (western) and Bol'shoi Kizil (eastern) flysch formations. Limestone conglomerates and the Biyagoda mixtite underlying the Bol'shoi Kizil Formation are genetically related to the latter and should be included into the Zilair Group as well (Fig. 6).

In the Western Magnitogorsk zone, the lower boundary of the Zilair Group is proposed to be placed at the base of clastic deposits directly overlying Frasnian cherts and mudstones, like it is accepted in the Voznesensk–Sakmara zone. Thus, it will be within the *triangularis* zone in southern areas and at the base of the *linguiformis* Zone of the conodont standard in other regions. In the greater part of Magnitogorsk megazone, the lower part of the Zilair Group is represented by the Sakmara Formation, whereas in the southern province and on the north of the northern province of the Western Magnitogorsk zone, it corresponds to the Bol'shoi Kizil Formation. The Karantau Formation is a component of the Prisakmara Formation, and the term becomes obsolete.

Lower boundaries of the Prisakmara and Bol'shoi Kizil formations are not synchronous because their clastic material is derived from different provenances, which evolved independently.

Formation of mixtites and conglomerates embedded in Upper Devonian deep-sea facies of the eastern flank of the Southern Urals and Mugodzhary was almost synchronous, corresponding to the middle part of the conodont *triangularis* Zone, though in places it lasted until the beginning of the *crepida* Zone. Their origin was presumably related to the sea-level drop, which reached its maximum in the middle part of the *triangularis* Zone. Differences in mixtite compositions observable in the Magnitogorsk megazone are determined by the state of their source area (island arc). On the north, where volcanoes were active, volcanogenic material was delivered to the basin, whereas on the south, where active volcanoes were missing, limestone buildups around the volcanic islands were subjected to erosion.

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Ammonite-Based Correlation of the Lower and Middle (*Panderi* Zone) Volgian Substages with the Tithonian Stage

M. A. Rogov

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 119017 Russia

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Abstract—Based on stratigraphic distribution of ammonite genera in common for the Volgian and Tithonian stages, a new correlation scheme of the lower–middle Volgian with the Tithonian is proposed. The lower Volgian Substage corresponds to the lower Tithonian Substage coupled with the middle Tithonian *semiforme* Zone at least. The *panderi* Zone is correlative with the interval from the middle Tithonian *fallauxi* Zone to the upper part of the upper Tithonian *microcanthum* Zone. The absence of ammonite genera in common hinders the direct correlation between higher levels of the Volgian and Tithonian stages. Peculiarities in Boreal–Tethyan migrations of ammonites during the early–middle Volgian are discussed. *Lingulaticeras blaschkei*, a species very important for correlation of the *panderi* Zone with the Mediterranean zonation is described and figures of some other ammonite species, such as *Sutneria asema* and *S.* cf. *eugyra*, are given for the first time.

Key words: ammonites, Boreal–Tethyan correlation, Volgian, Tithonian.

Since the 1960s, some specialists in stratigraphy of the Jurassic–Cretaceous boundary beds (e.g., R. Casey, A. Zeiss) called in question the equality of ranges of the Volgian and Tithonian stages. During last decades, new schemes based on mixed ammonite assemblages with Boreal and Mediterranean elements were proposed for correlation of these stratigraphic units. According to the Interdepartmental Stratigraphic Committee resolution, the Tithonian Stage is now considered as representing a stratigraphic equivalent of the lower–middle Volgian substages with the lower Volgian Substage corresponding to the lower–middle Tithonian [1]. Stratigraphic distribution of perisphinctids occurring both in the Boreal and Submediterranean realms is consistent with such a correlation.

Nevertheless, new data obtained during the last three decades on distribution of Submediterranean aspidoceratids, oppeliids, and haploceratids in the lower and middle Volgian substages of the Russian plate, which were virtually not taken into account until recently, allowed a new variant of the Volgian–Tithonian correlation to be proposed [2]. The availability of several different correlation schemes based on ammonoid succession implies the necessity to analyze all the data on co-occurrence of Subboreal and Submediterranean ammonite genera.

AMMONITE GENERA IN COMMON FOR THE VOLGIAN AND TITHONIAN STAGES: A REVIEW

In 1881, S.N. Nikitin [3] proposed the name *Volgian Formation* (Stage subsequently) for the post-Kimmeridgian sediments of the Russian plate because of a high provincialism of their ammonite faunas and impossibility of tracing in Russia the zonal units previously proposed for the Tithonian and Portlandian stages. Slightly later on, some ammonite species known from the Tithonian were found, however, in sediments, which are now referred to the lower Volgian Substage. For instance, Semenov [4] described *Aspidoceras* sp. from the Orenburg oblast. The examination of the Semenov's collection stored at the Chair of Historical Geology of the St. Petersburg University (no. 95) revealed that this form belongs to the lower Volgian species *Anaspidoceras neoburgense* (Oppel). The representative of the same species was likely described by Sokolov ([5], p. 23) as *Aspidoceras* sp. a few years later.[1] Soon after, Rozanov, ([6], p. 29) mentioned a fragment of ammonite "characterized by the Tithonian habitus and close probably to *Hoplites callisto* in sediments referred recently to the *virgatus* Zone and Subzone of the middle Volgian Substage. Unfortunately, this ammonite was not figured and no similar forms were ever recorded in the *virgatus* Zone.

Approximately at that time, first data on the occurrence of Boreal and Subboreal ammonites in areas adjacent to the Russian plate appeared. Par example, Abel [7] mentioned *Ammonites virgatus* from Lower Austria and Vetters ([8], Plate 22, fig. 5) figured *Perisphinctes* cf. *nikitini*. The last form was subsequently revised and described as a new species *Isterites austriacus* [9]. *Dor-*

[1] In his description, Sokolov paid attention to the absence of any tubercles in this form, which feature characterizes *A. neoburgense*.

soplanites panderi from the North Caucasus was also described and figured ([10], Figs. 1, 2). Simultaneously, Renz [11] reported on the occurrence of *Dorsoplanites dorsoplanus* and *Lomonossovella lomonossovi* in the North Caucasus, although he did not reproduce their figures.

In the 1930s, ammonites characteristic of the West European Jurassic were mentioned from Volgian sections in several works dedicated to geology of the Volga region. N.T. Zonov was the first to pay attention (without mentioning of particular sections) to the occurrence of a peculiar horizon with oppeliids within the Kimmeridgian–Volgian boundary beds of the Volga region. He emphasized ([12], p. 39) that "...of significant interest is the presence of poorly preserved forms similar to both some lower Tithonian *Ochetoceras* and upper Kimmeridgian *Haploceras*, for instance, to *H. (Glochiceras)* ex gr. *fialar*, in the uppermost layers with *Aulacostephanus*. Higher, these beds with *Glochiceras(?)–Ochetoceras* are defined as an autonomous marly unit lacking any signs of *Aulacostephanus*... The belonging of these beds to the *Oppelia (Ochetoceras) zio* Subzone, i.e., to the basal part of the *O. steraspis* Zone is highly probable". In the Gor'kii oblast and Chuvash Republic, Gerasimov and Kazakov [13] noted the occurrence of a peculiar horizon enclosing large bun-shaped concretions with *Oppelia* sp. and *Perisphinctes* sp. above the last finds of *Aulacostephanus*. Similar concretions are known in the Nizhnii Novgorod oblast from the uppermost Kimmeridgian (Isady locality) and Volgian (Murzitsy locality). At the same time, in the Chuvash Republic they are of the early Volgian age and contain either *Paralingulaticeras efimovi* (Rogov) and *Ilowaiskya* cf. *klimovi* (Ilov.) (Poretskoe Village, Sura River), or an older assemblage with *Neochetoceras steraspis* and *Lingulaticeras solenoides* (Polevye Bikshiki Village). A similar horizon with "rare slightly deformed ammonites referred to the *Ochetoceras–Haploceras (Glochiceras)* group" above the last finds of *Aulacostephanus* was also registered in the Tatar Republic [14].

Slightly later on, first figures of lower Volgian *Anaspidoceras neoburgense* (Oppel), which previously were only described or mentioned in the faunal lists, were reproduced by Ilovaiskii and Florenskii ([15], Plate XXIII, figs. 42, 42a). Soon after, this species was found in the Moscow outskirts and in the Ul'yanovsk oblast of the Volga region [16]. In addition, Sazonov [17] showed the occurrence of *Gravesia* in the lower Volgian Substage. This species previously established in England [18] grounded definition of the *gravesiana* Zone in the Russian plate [19]. The species from the Russian plate has not been figured, however, and is recorded only in the Gorodishche section (the upper part of the *klimovi* Zone), where ammonites are poorly preserved. Therefore, despite a wide distribution of *gravesiana* species that is characteristic of the lower Tithonian, Volgian, and Portlandian (sensu gallico), *Ilowaiskya klimovi* (Ilov.) is more suitable for being used as index species of the basal Volgian zone.[2] The species is common in sediments of the Russian plate, where it occurs beginning from the base of the Volgian Stage. Simultaneously, the transition from *Sarmatisphinctes fallax* (Ilow.) to *Ilowaiskya klimovi* (Ilov.) is very gradual and, if ammonites are poorly preserved, the boundary between Kimmeridgian and Volgian stages can be established more confidently based on the last occurrence of *Aulacostephanus*.

In their work dedicated to stratigraphy of the upper Upper Jurassic succession and relevant ammonites of southern Germany, Berckhemer and Hölder ([20], p. 58, Plate 14, fig. 68) described a species among ammonites from the upper Kimmeridgian–lower Tithonian referred it to the Boreal genus *Pavlovia (?Acuticostites)*. This form differing from typical *Acuticostites* in ornamentation patterns at early ontogenetic stages and in stratigraphic position should be attributed to the lower Tithonian genus *Berkhemeria* Schweigert et Zeiss, 1998.

Soon after, Mikhailov [21] demonstrated the occurrence of *Glochiceras* and *Neochetoceras* in the *klimovi* Zone of the Ul'yanovsk Volga region. In addition, he described a form from the *sokolovi* Zone and referred it, in the open nomenclature, to the species *Franconites vimineus* known from southern Germany ([22], p. 56, Plate 11, fig. 1).

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Works of the 1970s brought new information on joint occurrence of Tithonian and Volgian ammonites. A group of specialists who studied in detail the Gorodishche section under guidance of M.S. Mesezhnikov substantially augmented our knowledge on composition and stratigraphic distribution of Mediterranean elements in ammonite faunas from the lower–middle Volgian [39]. *Glochioceras (Paralingulaticeras)* was reported from *klimovi* and *sokolovi* zones, and *Neochetoceras* from the entire lower Volgian Substage, while *Sutneria* and *Haploceras* were found to occur up to the middle part of the lower Volgian *panderi* Subzone. Later on, *Pseudolissoceras* forms were found in the *panderi* Zone [40] and the range of *Haploceras* was extended in the North Caspian region up to the upper boundary of this zone [41]. Naturally, such a diverse assemblage of Tithonian ammonites found in the Volgian Stage attracted the researchers' attention and brought to idea that "*Glochiceras* and *Haploceras* found in the lower part of the *panderi* Zone enable a more confident correlation of the Gorodishche and Neuburg sections" ([42], p. 101). Later on, Mesezhnikov [41] proposed the first variant of that correlation. Unfortunately, the correlation was based on haploceratids only, and this resulted in assumption that upper boundaries of *bavaricum* and *panderi* zones coincide because Tethyan ammonites disappear at this level.[5]

Abundant data on Boreal ammonites from the Tithonian of West and East Europe and on some Submediterranean forms from the Volgian sediments of Poland appeared at the beginning of the 1970s. The Subboreal *Ilowaiskya* forms were found in Hungary [43], and *Zaraiskites* was detected in the upper Tithonian of the Polish Carpathians [44] and Bulgaria [45]. Figured ammonites identified as *Pavlovia iatriensis* Ilov. [46] have been reported from the upper Tithonian of Austria, where they occur together with *Pseudovirgatites*.[6] It should be noted that a similar assemblage from the Kletnice Beds includes mainly specimens collected by Vetters [8], positions of which in the section have not been indicated. Later on, it was established that they occur in the interval from the middle Tithonian *ponti* Zone to the upper Tithonian *simplisphinctes* Zone.

Kutek and Zeiss published several articles dealing with stratigraphy of the Volgian Stage in the Brzostówka section near Tomaszów Mazowiecki (Poland). They described several new *Pseudovirgatites* species from this section and a presumable *Lemencia* from the uppermost part of the lower Volgian Substage. In addition, they mentioned several *Isterites* species from the uppermost lower Volgian Substage of the Neuburg section and new species *Isterites mazoviensis* from the *scythicus* Zone. Precisely these finds, *Zaraiskites* forms from the upper Tithonian, on the one hand, and *Pseudovirgatites* and *Isterites* forms from the lower–middle Volgian, on the other, served as a basis for correlation of the lower Volgian Substage with the lower and middle Tithonian [48]. Soon after that, Malinowska [49] described the early Tithonian ammonites and paleobiogeography of the extra-Carpathian part of Poland. The described mixed assemblage of Subboreal *Ilowaiskya* and Submediterranean *Usseliceras* and *Subplanites* from the lower Volgian sediments also includes oppeliids and *Sutneria.* Subsequently, Schweigert [51] revised determinations of Submediterranean ammonites and attributed *Usseliceras* forms of Malinowskaya to the genus *Pseudovirgatites.*

After the detailed subdivision of the *scythicus* Zone in Poland, where two subzones and four faunal horizons were defined, Kutek [52] specified correlation between the Volgian and Tithonian stages. For instance, ammonites found in Bulgaria together with calpionellids [45] and appearing in the upper part of the *microcanthum* Zone were revised and determined as *Zaraiskites regularis* Kutek. This species characterizes a synonymous faunal horizon within the *zaraiskites* Subzone of the Volgian Stage in Poland, and Kutek correlated this horizon with an upper part of the *microcanthum* Zone. Simultaneously, Kutek and Zeiss [53] who studied drill cores established distribution range of *Neochetoceras* in Poland and concluded that the lower Tithonian should be correlated with the lower Volgian *klimovi* and *sokolovi* zones. Soon, *Neochetoceras*, *Haploceratina*, and *Sutneria* from the lower Volgian Stage of Poland were figured [28]. It should be noted that revision of earlier data led to conclusion that *Sutneria* and *Haploceratina* are missing from sediments younger than the *sokolovi* Zone.

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Schweigert [35] described some Boreal and Subboreal ammonites from the upper Kimmeridgian sediments (the *beckeri* Zone) of southern Germany, among them *Eosphinctoceras magnum* Mesezhn., the index species of the basal zone of the Volgian Stage in the sub-Polar Urals. With due regard to this species, he correlated the *magnum* Zone with the upper part of *autis-*

[5] These were haploceratids proper, as I think, because Submediterranean *Isterites* exist above the *bavaricum* Zone in the Neuburg section.

[6] According to Mesezhnikov [50], the ammonite specimen figured by A. Zeiss is indeterminable and cannot be attributed to *Pavlovia iatriensis*.

siodorensis Zone. However, *Aulacostephanus autissiodorensis* (Cotteau) is unknown in the sub-Polar Urals, and the synonymous zone is established there based on indirect evidence. The *magnum* Zone, at least partly, is probably of the Kimmeridgian age, because *Gravesia lafauriana* of Hantzpergue [54] (=*Gravesia polypleura*, Zakaharov and Mesezhnikov [55], Plate 1, fig. 1) was found at this level in the sub-Polar Urals. However, a distinctly Tithonian *Gravesia* form (*?Gravesia gravesiana* [54] = *Gravesia* sp.[55], 1974, Plate 1, fig. 2) is reported to occur in the same zone, and both finds of *Gravesia* are confined to the lower part of the zone. Therefore, the Kimmeridgian–Volgian boundary in the sub-Polar Urals can be placed preliminarily within the *magnum* Zone. On the other hand, the oppeliid succession suggests that the base of the Volgian Stage in central Russia corresponds to the base of the Tithonian Stage in Germany. The terminal faunal horizon of the upper Kimmeridgian Substage in Germany and the upper part of the *fallax* Zone in the Russian plate contain one and the same form *?Neochetoceras frodingense* (Berck. et Hölder) that replaces in both areas *N. rebouletianum* and *N. zlatarskii* with coarse ornamentation ([56], Plate 1, fig. 5).

Recently, two early Tithonian species described from the Neuburg section [57] were attributed to *Sarmatisphinctes* and *Dorsoplanites* [58]. Unfortunately, their attribution to Boreal genera is substantiated insufficiently, because both forms are represented by holotypes only and their ornamentation in ontogenesis is unknown. The idea to regard *Dorsoplanites lumbricarius* (Schneid) as a possible ancestor of *D. panderi* (Orb.) [57] is also poorly substantiated, because these species are separated by a stratigraphic gap and characteristic of different basins.

AMMONITE-BASED VARIANTS OF CORRELATION BETWEEN VOLGIAN AND TITHONIAN STAGES

Relationships between the Volgian and Tithonian stages are mainly interpreted based on ammonites in common for both stratigraphic units. Joint finds of Tethyan ammonites and buchiids play a significant role in correlation of upper Volgian and younger sediments.

Inferences of researchers used to be based on different ammonite groups. For instance, Kutek and Zeiss [59] take into consideration mainly the distribution of perisphinctids, such as *Zaraiskites, Pseudovirgatites, Pavlovia*, and *Isterites*. Only recently, they used data on distribution of oppeliids *Neochetoceras steraspis* (Oppel) and *N. mucronatum* (Berck. et Hölder) to substantiate correlation of the lower Tithonian with the Volgian *klimovi* and *sokolovi* zones [60]. It should be noted that the group of variable perisphinctids consists of several parallel phylogenetic lineages, interpretation of which considerably affects correlation. Par example, if ammonites from the Neuburg section originally identified as *Zaraiskites* cf. *zaraiskensis* (Mich.) [37] are taken for true *Zaraiskites zaraiskensis* [61], or for homeomorphs only [38], the resultant biostratigraphic schemes turn out to be different by a substage. On the other hand, if we take into consideration a poor preservation of late Tithonian *Zaraiskites* and admit, following Kutek and Zeiss, that descendant *Isterites* with *Virgatites*-like ribs existed in the Submediterranean Province, we should question whether they are true *Zaraiskites* or not? Anyway, neither ornamentation patterns nor peculiarities of their septal suture are known.

It should be noted that correlation of the lower Volgian Substage with the lower–middle Tithonian does not contradict in principle to assumable equivalency of the Volgian and Tithonian stages. Mikhailov [62] accepted precisely that variant of correlation, which allowed him to admit correspondence of the Volgian and Tithonian stages. In their early correlation scheme, Kutek and Zeiss ([9], p. 513, Table 1) also admitted equality of these stages, despite the fact that they correlated the upper Tithonian with the lower (partly), middle, and upper Volgian substages.

On the other hand, Mesezhnikov [63] substantiated his interpretations based on Tethyan elements occurring in the Volgian Stage of the Russian plate. He proposed to correlate upper boundaries of the *bavaricum* and *panderi* zones, assuming that haploceratins disappear synchronously in the Neuburg section and Russian plate. In my opinion, the disappearance of that ammonite group in two separate basins, while it continued to exist southerly, does not necessarily indicate synchronism of the events. In addition, the impoverished taxonomic composition of ammonites in the upper part of the Neuburg Formation is likely connected with desalination, whereas signs of similar event cannot be observed across the boundary between *panderi* and *virgatus* zones.

In 2001, I undertook a preliminary attempt to compile all the available data on ammonites occurring in both the Volgian and Tithonian stages [64], but the detailed analysis of their distribution was out of the scope of that work.

STRATIGRAPHIC AND GEOGRAPHIC RANGES OF AMMONITE GENERA FROM VOLGIAN AND TITHONIAN STAGES

A correct correlation of the Tithonian and Volgian stages should be based on firmly established stratigraphic ranges of ammonites occurring in both units. Unfortunately, many ammonite taxa important for solution of the problem have not been figured or described, being mentioned only among others. In discussion below, I consider therefore the probable erroneous identifications of the forms in question. Nevertheless, it seems reasonable to assume that ammonite taxa listed in publications are correct at the generic level at least. We may also assume that stratigraphic ranges of Tethyan ammonites in the Russian plate and adjacent areas of West Europe are comparable, although they

can be different in remote areas. At present, we know over ten ammonite genera occurring in both the Volgian and Tithonian stages and considered below (in all cases, unless it is indicated otherwise, the Volgian haploceratins from the Russian plate are those from the Gorodishche and Polevye Bikshiki sections, the Batyrevskii area of the Chuvash Republic).

1. *Pseudolissoceras.* In European Russia, representatives of this genus are known only from the *panderi* Zone [40].[7] I found disputable specimens of the genus in the middle part of the *pseudoscythica* Zone (*neoburgense* faunal horizon) ([65], Plate, fig. 10). In Hungary and the Polish Carpathians, the genus occurs up to the middle Tithonian *fallauxi* Zone [66]. In Romania [67] and Hungary [43], its representatives disappear in the *fallauxi* Zone. They occur up to the top of the *fallauxi* Zone also in Spain [68] and to the middle part of this zone in Italy [69]. Thus, it is reasonable to conclude that the *bavaricum* Zone of the Neuburg section (and the lower part of the *panderi* Zone at least) is not older than the *fallauxi* Zone [61], being incompletely correlative to the middle Tithonian, as is frequently accepted (I mean correlation of the *bavaricum* Zone with the entire middle Tithonian [70]).

The stratigraphic range of the genus in southerly areas is substantially wider. It occurs in the lower part of the upper Tithonian in Algeria and Iraq [71] and up to the lower part of the Berriasian (analogues of the *euxinus* Zone) in Tunisia [72], as it was noted first by Imlay [73].[8] This distribution peculiarity was used by Jeletzky [61] for age determination of the Neuburg Formation.

In Central and South America, *Pseudolissoceras* likely occurs in the middle Tithonian only (the middle Tithonian was originally established based on the taxon occurrence precisely in this region). Burckhardt [75] described several *Pseudolissoceras* species (the type species inclusive) from Mexico and later on, discussing the taxon occurrence in the Mexican middle Tithonian, Verma and Westermann [76] correlated partly the *zaraiskensis* Zone with this level also (I am of the same opinion). Imlay [77] reported, however, on the *Pseudolissoceras* occurrence in the lower upper Tithonian of Mexico, where its species (*P. zitteli* included) coexist with *Durganites* and *Buchia mosquensis* (!!).[9] In Argentine, *Pseudolissoceras* is known from the lower part of the middle Tithonian [78].

In the Far East, *Pseudolissoceras* occurs in the *zitteli* Zone (=*primoryense* Zone after Khudolei [79]).[10] Sei and Kalacheva [80] correlate this zone with the synonymous unit of Argentine, which corresponds approximately to the *semiforme* Zone [82]. In opinion of Khudolei [79], it corresponds to the *bavaricum* Zone of the Franconian Alb. Based on finds of *Buchia mosquensis* and *B. rugosa* in the same layers, Sei and Kalacheva [83] assumed that these zones are correlative with the lower to middle (partly) Volgian sudstages. It should be noted that the mentioned *Buchia* species continue to occur also in the *panderi* Zone of the Russian plate [84] and *Buchia mosquensis* is abundant in the *Epivirgatites nikitini* Zone of East Siberia [85]. In opinion of Parent [86] who revised the genus in question, none of figured *Pseudolissoceras* forms is known even from the upper part of the *fallauxi* Zone. Thus, *Pseudolissoceras* forms do not occur in Europe above the *fallauxi* Zone, and their presense in the middle Volgian *panderi* Zone indicates that the last unit corresponds, at least partly, to the *fallauxi* Zone.

2. *Haploceras.* This genus is of little use for reliable correlation, because its shells are poorly preserved in the Russian plate, being undeterminable at the species level (the *Haploceras* sp. in lists),[11] and the genus proper occurs almost worldwide from the lower Kimmeridgian to the uppermost Berriasian. In addition, most of European species of the genus represent probably a single form characterized by polymorphism [87]. In the Gorodishche section, *Haploceras* sp. occurs at least from the upper part of the *klimovi* Zone ([65], Plate, fig. 11) to the *panderi* Zone of the middle Volgian Substage [88]. In addition to the Gorodishche section, *Haploceras* form are known from the upper part of the *pnaderi* Zone of the North Caspian region [41].

3. *Lingulaticeras.* In the Russian plate, *Lingulaticeras* s. l. are recorded in the interval from the upper Kimmeridgian *autissidorensis* Zone to the middle Volgian *panderi* Zone [89]. In addition, Gerasimov's collection (PIN no. 4861) includes one undoubted specimen of *Lingulaticeras blaschkei* (Cecca et Enay) from the *panderi* Zone found in the Kimry outskirts, the Tver oblast, and V.V. Mitta donated me several specimens of this species from the *panderi* Zone of the Chuvash Republic.[12] When defining the *Glochiceras* subgenera (individual genera in this work), Ziegler [91] demonstrated importance of the aperture structure (presence or absence of hood, incline of the apertural margin, and other features). The aperture is rarely preserved in middle Tithonian *Glochiceras* forms, only one of which was figured by Blaschke ([92], Plate 1, fig. 7). Judging from its structure, the *blaschkei* form can confidently be attributed to the genus *Lingulaticeras*, like the problematic ammonite forms, which were recently found in the *neoburgense* faunal horizon of the Gorodishche

[7] According to oral communication of A.G. Olfer'ev, data on ammonites in the scheme were summarized by M.S. Mesezhnikov. Unfortunately, no *Pseudolissoceras* forms are found so far in the Mesezhnikov's collection stored at the VNIGRI.

[8] Howarth [74] had cast doubts on the occurrence of post-middle Tithonian *Pseudolissoceras* forms both in Tunisia and Iraq because they were never figured.

[9] The forms have not been figured.

[10] Sei and Kalacheva [81] consider *Primorytes primoryense* Chud. 1960, as a synonym of *Pseudolissoceras zitteli*, thus changing the zone index for *zitteli*.

[11] Upper Kimmeridgian ammonites from the Gorodishche section determined as *Haploceras* most likely belong to *Neochetoceras* ex gr. *subnudatum*.

[12] Previously, I attributed these forms to *Glochiceras* aff. *contractum* [90].

section. In Germany, *Glochiceras* s. l. are undoubtedly known from sediments, which are younger than the lower part of the *bavaricum* Zone, and in the Neuburg section, they occur in Bed 22 [93]. The same bed yields *Virgatosimoceras* cf. *albertinum*, the index species of the *darwini/albertinum* Zone, the upper one of the lower Tithonian [93, 94]. In Poland, probable *Glochiceras* s. l. are mentioned form the *klimovi* and *sokolovi* zones [28], although they are not figured, like in Russia. In France and Italy, *Glochiceras* s. l. is characteristic of the middle Tithonian *fallauxi* Zone [95]. Moreover, *Lingulaticeras blaschkei* is described from the Ardéche section in France. Unfortunately, the exact stratigraphic positions of *L. blaschkei* in Czechia (Stramberg) and the North Caucasus are not indicated. *"Oppelia strambergensis"* mentioned (but not figured) from the middle Tithonian of the Crimea [96] probably belongs to the genus *Lingulaticeras*. Poorly preserved *Glochiceras* s. l. are known from the middle Tithonian of Romania [97]. In Madagascar, *Lingulaticeras* is found in the middle Tithonian *kobelli* Zone [98]. In addition, it cannot be ruled out that some ammonites of Madagascar, which are referred to the genus *Hildoglochiceras*, belong in fact to *Lingulaticeras* forms (e.g., *"Hildoglochiceras" nudum* Collignon, 1960). Specimens of *Lingulaticeras umbilicocrenatum* Collignon described from Madagascar are also known from the upper part of the middle Tithonian in the Far East [81]. Until recently, ammonites of the last region were classed with *?Glochiceras jollyi* Oppel. However, shells of *Ammonites jollyanus* ([99], p. 271, Plate 75, fig. 4) are up to10 cm across (i.e., twice larger than the largest known *Glochiceras* specimens) and have no lateral furrow. Moreover, their septal suture is of peculiar shape different from that of *Glochiceras*, and the aperture structure is unknown. In opinion of H. Schreier (oral communication), the original specimen from the Oppel's collection (SMNS, Inv. No. 1872 XV 4) should probably be referred to Triassic Gymnitidae such as *Ammonites lamarcki* Oppel, 1863, a type species of the genus *Anagymnites* Hyatt, 1900. This is consistent with the viewpoint of specialists on Triassic ammonites, beginning from Diener [100] who consider this species as the Triassic one. Therefore, I attribute the forms described by Sei and Kalacheva [81] and by Collignon [98] to species *L. umbilicocrenatum* [101]. Simultaneously, *Glochiceras* forms from the Primor'e bear some peculiar features untypical of *L. umbilicocrenatum* and, in the opinion of Sei and Kalacheva (pers. comm.), they belong most likely to a new species. *Glochiceras* forms mentioned from the middle Tithonian of Iraq [102] should probably be referred to the genus *Hildoglochiceras*. *Glochiceras* specimens are also known from Cuba (*Glochiceras (?Lingulaticeras)* sp. in [103], p. 291, Plate 2, figs. 1, 2, 4, 7–9). These ammonites are virtually identical to *L. umbilicocrenatum* (Collignon) from the Primor'e. In Antarctica [104], *Glochiceras* similar to *Lingulaticeras* (=Oppelid gen. nov., Plate III, fig. g) is found in the lower Tithonian. Morphologically similar ammonites of the genus *Hildoglochiceras* are known from the middle Tithonian only. Enay [105] considered these forms as confined to the upper Tithonian, but new data suggest their middle Tithonian age [106]. At the same time, the late Tithonian *"Rapidoceras"* and *Hildoglochiceras (Salinites)* from Mexico and Cuba have the distinctive ornamentation in the upper part of the whorl.[13] In addition, *Hildoglochiceras (Salinites)* forms, which occur in Mexico up to the lower part of the upper Berriasian Substage, have the well-developed keel. Therefore, the middle Tithonian *Lingulaticeras* from Cuba and the Primor'e can hardly be referred to the subgenus *Salinites*, as it was done by Myczyński [109]. In Ardéche section (France), *Lingulaticeras blaschkei* (Cecca et Enay) characterizes a relatively narrow interval within the middle part of the *fallauxi* Zone (Beds 8a, 10) reflecting the second and third episodes of the strengthened Mediterranean influence during this period [110].

4. *Paralingulaticeras*. Representatives of this genus were found in the *klimovi* and *sokolovi* zones of central Russia [111]. In addition, the Mesezhnikov's collection (VNIGRI) includes *Paralingulaticeras* with labels indicating the "sokolovi" Zone. Unfortunately, it is unknown, whether these ammonites characterize the entire zone or a part of it. Judging from the light coloration of the host rock, they are from the lowermost part of the zone in question, which encloses also *Ilowaiskya sokolovi* Ilow.

All the *Paralingulaticeras* specimens from the Russian platform can probably be attributed to one very variable species *Paralingulaticeras efimovi* (Plate, fig. 7). This species differs from European *Paralingulaticeras* forms in smaller dimensions and absence of developed ventrolateral tubercules. Small dimensions of *Paralingulaticeras* representatives from the Russian platform and ornamentation characteristic of inner whorls of European *L. lithographicum* suggest neoteny in their development. Similar populations were characteristic of Boreal ammonites that penetrated into the Submediterranean Province [112].

First finds of subgenus *Paralingulaticeras* are reported from the *eudoxus* Zone [113], although these early forms are of peculiar appearance. In the Franconian Alb [36], rare *Paralingulaticeras* specimens appear beginning from the base of the *lithographicum* Subzone in the upper part of the Solnhofer Formation to become abundant in the overlying Mörnsheimer Formation.[14] In the Schwabian Alb, rare *G. lithographi-*

[13]The former genus was describred without diagnosis and indication of the type species [107] among Cuban ammonites previously referred to the genus *Haploceras* [108].

[14]The *lithographicum* Subzone is defined in the Franconian Alb by Berthel and Schairer ([114], p. 12): "Das untere Untertithon umfät die Zone des *Hybonoticeras hybonotum*; ihren oberen Abschnitt kann man als Subzone des *Glochiceras lithographicum* (Obere Solnhofer Plattenkalke und Mörnsheimer Schichten) ausschieden". Its base is placed at the appearance level of *G. lithographicum*.

cum occur in the *riedlingensis* faunal horizon of the *hybonotum* Zone [115]. According to G. Schweigert (pers. communication), this area is unfortunately lacking the level with abundant *P.* ex gr. *lithographicum* and, thus, the *efimovi* faunal horizon can be correlated with Schwabian biohorizons only arbitrarily. These ammonites occur in the lower Tithonian *hybonotum* Zone and its analogues in Spain [116], Italy [117], Poland [66], Bulgaria [118], Czechia [119], Romania [120], France [121], and Albania [122]. Noteworthy is that in Poland they are unknown from the *klimovi* Zone, while the *sokolovi* Zone encloses forms very similar to coarsely costate *P. lithographicum* (=*Ochetoceras* or *Taramelliceras*, [28], Plate 31, fig. 7). *Paralingulaticeras* is characteristic almost exclusively of Europe, where it never occurs above the *hybonotum* Zone. Beyond Europe, one locality of this subgenus is known in Madagascar [123], and one specimen (not figured) is reported from Antarctica [124].

Although *Paralingulaticeras efimovi* is unknown outside the Volga region, it can be used for correlation, because its appearance was concurrent to wide distribution of other *Paralingulaticeras* forms in West Europe. If events of *Paralingulaticeras* disappearance were synchronous in different basins, then the lower part of the *sokolovi* Zone in the Russian plate should be correlated with the upper part of the *hybonotum* Zone. The *efimovi* faunal horizon of Central Russia corresponds to a greater part of Mörnsheimer Beds of the Franconian Alb, which yield abundant *Paralingulaticeras* [36], and to a portion of the *laisackerensis* faunal horizon (and, probably, to cf. *eystettense* faunal horizon) of the Schwabian Alb [65]. It cannot be also ruled out that the base of the *efimovi* faunal horizon corresponds to the *riedlingensis* horizon of Schwabia, where first, although not numerous *Paralingulaticeras* specimens appear.

5. *Neochetoceras.* In the Gorodishche section, *N.* cf. *steraspis* are persistently occurring in the *klimovi* Zone [125]. Nevertheless, it seems quite probable that some *Metahaploceras* could be taken for *Neochetoceras.* According to my observations, *N. steraspis* replaces species *N. fridigense* and *N.* ex gr. *subnudatum* occurring in the uppermost Kimmeridgian Stage of the Volga region. In the Gorodishche section, this species is common in the lower part of the *klimovi* Zone (Beds with *Lingulaticeras solenoides* and *Neochetoceras steraspis*) but rare in the overlying *efimovi* faunal horizon [65]. Recent data indicate the *Neochetoceras* occurrence in the higher *pseudoscythica* Zone [40]; the Mesezhnikov's collection (VNIGRI) includes ammonites from the *sokolovi* Zone close to *Neochetoceras.* Because of poor preservation and gradual morphological changes of ammonites in majority of the examined sections, it is difficult to trace the transition from Kimmeridgian to Volgian *Neochetoceras* species. For instance, the Kimmeridgian–Tithonian boundary layers in the Gorodishche section include an interval approximately 0.3 m thick, where *Neochetoceras* specimens are indeterminable at the species level and other stratigraphically important ammonites are missing. *Neochetoceras* forms occur beginning from the second half of the Kimmeridgian to the beginning of the middle Tithonian in Germany [126] and also in younger sediments (up to the middle Tithonian *fallauxi* Zone) in Spain [127], France [128], and Italy [129]. A similar distribution of the genus is probably characteristic of Mexico [130]. The youngest *Neochetoceras* forms are known from the *microcanthum* Zone of Hungary [131]. In Poland, *N. steraspis* occurs in the *klimovi* Zone and close *N. mucronatum* in the upper part of the *klimovi* and in the *sokolovi* zones [28]. In the North Caucasus, *N. praecursor* is known from the basal part of the lower Tithonian [132]. This species characterizes as well the *eigeltingense* faunal horizon, the lowermost Tithonian unit of southern Germany [133]. In the Russian plate, analogues of this level are not established so far. In the Gorodishche section, *N.* cf. *steraspis* appears abruptly at the base of the *klimovi* Zone. Recently, *Neochetoceras* sp. of poor preservation was reported from the *hybonotum* Zone of Antarctica ([124], fig. 7A). Representatives of the genus migrated to this region most likely from East Africa, where they are known for a long time [134]. In addition, *Neochetoceras* sp. was recently found in the Andes of Argentine: in the lower Tithonian *mendozanus* Zone [135] and in the middle Tithonian *proximus* Zone, an approximate analogue of the *fallauxi* Zone [136].

According to data on *Neochetoceras* distribution, the *pseudoscythica* Zone is, at least partially, not younger than the middle Tithonian *fallauxi* Zone. Beds with *Neochetoceras steraspis* and *Lingulaticeras solenoides* of central Russia approximately correspond to *rueppelianum* and *riedlingensis* faunal horizons and to unnamed unit between the *rueppelianum* and *eigeltingense* horizons of the Schwabian Alb. *Neochetoceras* ex gr. *steraspis* and *Lingulaticeras solenoides* occur at all these levels, while *Paralingulaticeras* forms are missing or scarce (G. Schweigert, pers. communication).

6. *Fontannesiella.* I found several *Fontannesiella* aff. *prolithographicum* specimens among ammonites from the *efimovi* faunal horizon of the Gorodishche section ([65], plate, fig. 5). Unlike typical *Fontannesiella*, these ammonites, as well as coexisting *Paralingulaticeras*, are lacking developed tubercles. *F. prolithographicum* associate usually with *Paralingulaticeras lithographicum* and probably represent macroconchs of the latter [137]. Finds of *F. prolithographicum* are known from the *hybonotum* Zone of the North Caucasus [138], Spain [127], Portugal [139], Ethiopia [140], southern Germany [141], southeastern France [142], and Sicily ([143], fig. 144), while forms determined in the open nomenclature are reported from Antarctica ([124], fig. 6C). The species never occurs above the *hybonotum* Zone, being reported from the upper Kimmerdgian *beckeri* Zone of southeastern France only [144]. Other *Fontannesiella* species, such as *F. valen-*

tina, are sometimes mentioned among ammonites from the higher stratigraphic level of the *darwini* Zone, where *Paralingulaticeras* is unknown [145].

7. *Sutneria.* This genus is widespread almost everywhere in the Kimmeridgian of the Northern Hemisphere and in the Tithonian of Europe. In the Russian plate, representatives of the genus span the interval from the upper Kimmeridgian to the *panderi* Zone [39], where Volgian forms are very scarce. The Beds with *Neochetoceras steraspis–Lingulaticeras solenoides* of the *klimovi* Zone contain *S.* cf. *eugyra* (Plate, fig. 4) in several sections. Similar *Sutneria* forms are also known from the *klimovi* Zone of Poland ([28], *Sutneria* cf. or aff. *bracheri*, Plate 31, figs. 1–4). *S. eugyra* is a species of the *hybonotum* Zone (*laisackerense* faunal horizon) of southern Germany [115] and of the lower Tithonian in Romania [146].

In the Gorodishche and Polevye Bikshiki sections, relatively abundant *Sutneria asema* occur only in the *neoburgense* faunal horizon. This species is recorded up to the *fallauxi* Zone in the Polish Carpathians [66] and Romania [147]. In Germany, *S. asema* is known from two lower faunal horizons of the Neuburg Formation [58], where they associate with *Pseudolissoceras bavaricum* and *Anaspidoceras neoburgense* [93]. *S. asema* was also found in the upper part of the lower Tithonian: in the *darwini* [148] and (in Germany) *vimineus* zones (A. Scherzinger, personal communication). Representatives of the genus *Sutneria* are also reported from Argentine, but their stratigraphic position remains unclear so far (H. Parent, pers. communication). Slavin ([149], Plate II, figs. 11–14) attributed ammonites from the lower Valanginian (=Berriasian) sequences of the Ukrainian Carpathians to *Eurynoticeras* aff. *asema* (= *Sutneria*). These forms are lacking costae at their sides, however, and have instead small constrictions in the lower part of the whorl. Such morphological peculiarities are untypical of *Sutneria* (e.g., [150], Plate III, figs. 8, 9; lectotype of *S. asema*). Unfortunately, Slavin did not illustrated the lobe line of these ammonites, which most likely belong to the genus *Ptychophylloceras*. Khalilov and Abdulkasumzade [151] meant probably these very forms, when they argued that *S. asema* ranges from the Kimmeridgian to Berriasian. The last *Sutneria* specimens are characteristic of the *semiforme* Zone or of the level that is not higher, at least, than the *fallauxi* Zone of the middle Tithonian. Consequently, the lower part of the *panderi* Zone is not younger than the *fallauxi* Zone.

8. *Pseudovirgatites.* In Russia, ammonites similar to *Pseudovirgatites* described from the upper part of the *pseudoscythica* Zone of platform areas in Poland [48] were figured by Mikhailov [22] under the name *Pectinates (Wheatleyites) arkelli.* Unfortunatelly, all specimens figured by him are fragmentary, and morphology of their internal whorls (more exactly, positioning of costae branching points) is unknown. *Pseudovirgatites* specimens are reported from both the Vetlyanka and Gorodishche sections (similar forms were described by Semenov [4] as *Perisphinctes capillaceus*). In my collection, there are specimens of *P. puschi* found in the *puschi* faunal horizon crowning the *pseudoscythica* Zone in the Gorodishche and Polevye Bikshiki sections ([65], Plate, fig. 13). One ammonite fragment with the high branching coefficient of costae from the *puschi* faunal horizon of the Gorodishche section can be identified as *P.* aff. *seorsus* (Plate, fig. 8). Well known in West Europe are *Pseudovirgatites* specimens (the species unknown from Poland and central Russia) occurring mainly in the upper Tithonian. They are described from France [121], Spain [152], the Polish Carpathians [66], and Hungary [43], where they occur up to the upper Tithonian. The less frequently mentioned are the middle Tithonian or stratigraphically uncertain *Pseudovirgatites* forms. They are reported from the middle Tithonian of Hungary ([153]; new species) and uppermost middle Tithonian of Czechia (Stramberg) [61] together with "ancient ammonites" of the *fallauxi* or *ponti* zones. Nowak [154] reported on *Pseudovirgatites* associated with *Semiformiceras fallauxi. Pseudovirgatites* is also known from the middle Tithonian of Romania [67]. In Austria, representatives of this genus are found together with *"Pavlovia iatrensis"* [46] (my comments to this species see above), although formerly they were described from the same localities together with *Semiformiceras semiforme* and *Uhligites lymani* [155] characteristic of the middle Tithonian. Relationships between *Pseudovirgatites* and *Zaraiskites* remain unclear so far. Some researchers [156] consider the former taxon as ancestor of the latter (I share this viewpoint) and see, following Ilovaiskii and Florenskii[15], the main difference between genera in different position of the costae branching points, which is low on internal and higher on external whorls of *Pseudovirgatites* and vice versa in *Zaraiskites* shells. Ornamentation patterns on external whorls of some *Zaraiskites quenstedti* are virtually identical to those of *Pseudovirgatites* forms.

Kutek and Zeiss considered *"Ilowaiskya" tenuicostata* (Mich.) as ancestral taxon of *Pseudovirgatites.* This species appears earlier than *Pseudovirgatites puschi.* In the Gorodishche section, its first representatives (Plate, fig. 9) are known from the upper part of the *neuburgense* faunal horizon. Consequently, the *puschi* and *neoburgense* faunal horizons can be considered as subdivisions of the *tenuicostata* Subzone. *"Ilowiaskya" tenuicostata* shows morphological features of both *Ilowaiskya* (absence of costae characteristic of *Virgatites*) and *Pseudovirgatites* (low position of costae branching point on internal whorls). This species begins most likely the *Pseudovirgatites* phylogenetic lineage and consequently belongs to this genus. Thus, *Pseudovirgatites* from the Volgian Stage definitely suggest only that the *tenuicostata* Subzone (at least) of the *pseudoscythica* Zone can be correlated with a part of the middle Tithonian. Discrimination of the *tenuicostata* Zone in the Russian plate is inconsistent with the

fact that corresponding beds yield true *Ilowaiskya* forms (*I. pseudoscythica* of the *neoburgense* faunal horizon) and can be ranked as a subzone only.

9. *Danubisphinctes.* In the European part of Russia, undoubted finds of these ammonites are unknown. Only the Polevye Bikshiki section yields rare and small coarsely ornamented ammonites, which can be attributed to this genus. In addition, *Perisphinctes (?Ilowaiskya)* sp. indet. (not figured) is suggested to resemble ([15], p. 107) *Pseudovirgatites (Danubisphinctes) palmatus subpalmatus.* Species of the genus *Danubisphinctes* (=*Isterites* after Kutek and Zeiss) occur in the *tenicostata* Subzone of the lower Volgian and in the *scythicus* Zone of the middle Volgian in Poland [157]. *Danubisphinctes* is widespread in the lower–middle Tithonian sediments, and its late coarsely ornamented forms (=*Isterites* auct.) are of interest for this work. In the Neuburg Formation, young representatives of the genus appear in the middle part of the Unterhausen Beds (Bed 60) and disappear in the lower part of the Oberhausen Beds [158]. The *palmatus* faunal horizon, the highest ammonite-containing level of the formation, yields *Danubisphinctes* species known from the *puschi* Zone of Poland. In Hungary, *?Isterites* sp. is reported from the *"Burckhardticeras"* Zone, which used to be correlated with the terminal *ponti* Zone of the middle Tithonian [43].[15] The same age is characteristic of *"Isterites"* from Spain [159]. *D. spurious* present in the upper part of the *tenuicostata* Zone of Poland and in the upper faunal horizon of the Neuburg Formation indicates a partial overlapping of these stratigraphic units.

10. *Ilowaiskya.* In the Russian plate, representatives of this genus occur only in the lower Volgian Substage as index species of all zones. A similar stratigraphic range is characteristic of them also in Poland, where *Ilowaiskya* s. l. is missing only from the *tenuicostata* Subzone. In southern Germany, the genus appears in the upper part of the Rennertshofener Formation (the *palatinum* Zone) [160]. *Ilowaiskya* cf. *pseudoscythica* (not figured) from the Neuburg Formation [161] occurs in the lower part of the Unterhausen Beds and disappears before the first occurrence level of *Isterites.* In Bed 22, the species associates with *Virgatosphinctes* cf. *albertinum*, the index species of synonymous zone correlated with the *darwini* Zone [94].[16] In Hungary, *Ilowaiskya* ex gr. *klimovi* (not figured) is found in the basal *hybonotum* Zone of the lower Tithonian [43]. Unfortunately, none of Tithonian *Ilowaiskya* forms is figured, except for *"Ilowaiskya* aff. *pavida juvenilis"* ([37], Plate 22, fig. 4; =*Sublithacoceras penicillatum* (Schneid) after Scherzinger and Schweigert [58]), and it is probable that researchers considered the morphologically close representatives of Mediterranean perisphinctids. For instance, Scherzinger and Schweigert [58] revised *Ilowaiskya* cf. *pseudoscythica* ([37], p. 117) and identified it as *Danubisphinctes. Ilowaiskya* forms are quite suitable for correlation of lower Volgian sections in different regions. Their distribution in the Tithonian Stage (provided that identifications are correct) suggests that the *pseudoscythica* Zone is correlative, at least partly, with the *darwini* Zone of the lower Tithonian.

11. *Zaraiskites.* In European Russia and Poland, *Zaraiskites* occurs in the lower part of the middle Volgian Substage. Ammonites from the Neuburg Formation first described as *Z.* cf. *zaraiskensis* [37] were subsequently attributed to other genera [38]. Single *Zaraiskites* specimens are known from the upper Tithonian (the age confirmed by *Crassicolaria*) of Bulgaria [45] and the Polish Carpathians [44], although preservation of these ammonites is not perfect. One more specimen is reported from upper Tithonian sediments of Austria [46]. Recently, Zeiss ([162], p. 62, Plate 14, fig. 2) figured a small ammonite fragment from Ernstbrunne determined as *Zarajskites*, but its poor preservation hinders identification even at the generic level. Kutek [52] attributed the specimen figured by Nowak [45] to *Z. regularis* Kutek, 1994, and correlated the synonymous faunal horizon of the *zaraiskensis* Subzone of Poland (the middle Volgian *scythicus* Zone) with the upper Tithonian *transitorius* Zone. As Kutek noted himself, *Zaraiskites* forms are rather highly variable. This peculiarity and a poor preservation of ammonites from Bulgaria are the obstacles for their confident identification at the species level. In any case, virgatotome ribbing is also characteristic of many Mediterranean middle–upper Tithonian ammonites, e.g., it is typical of internal whorls of *Danubisphinctes mutabilis* ([163], fig. 9). Nevertheless, if the genus identification is correct, it is highly probable that the upper part of the *panderi* Zone is correlative partially with the upper Tithonian *microcanthum* Zone. The *Zaraiskites* successions in Poland and the Volga River basin are likely similar. Anyway, combustible shales of the *panderi* Zone of the Volga region enclose *Z. regularis.*

12. *Anaspidoceras.* Ammonites from the *neoburgense* faunal horizon of the Gorodishche section, which are identified as *A. neoburgense* [164], have never been figured, but my collection includes determinable specimens ([65], plate, fig. 12). According to Cecca [165], *A. neoburgense* crowns the evolutionary lineage of the *Anaspidiceras* genus and is a single species lacking tubercles like specimens from the Gorodishche section. In distinction from ammonites of the last section, the form figured by Ilovaiskii and Florenskii [15] is intact and confidently identified. It seems that Sokolov ([5], p. 23) and Semenov ([4], p. 182) described *A. neoburgense* from the lower Volgian sediments of the Vetlyanka locality as *Aspidoceras* sp. (in any case, ornamentation patterns of the latter are identical to those of *A.*

[15]The Upper Jurassic genus *Burckhardticeras* Olóriz, 1978, is a homonym of the Lower Jurassic genus *Burckhardticeras* Flores Lopez, 1967 [51], and its name is given in quotes.

[16]In the recent work dedicated to the Neuburg section [58], this species is not mentioned, although the close level is indicated as yielding *V. broilii* [57] characteristic of the same zone.

neoburgense).[17] In the Polish Carpathians, the species in question occurs mainly in the *darwini* and *semiforme* zones [66] and appears again, after a significant stratigraphic gap, in the lower Berriasian *euxina* Zone ([166], Plate 2, fig. 7). In Germany, the species is characteristic of the *ciliata* Zone [58]. In Spain, *A. neoburgense* was also found in the basal Berriasian *jacobi/grandis (=euxinus)* Zone. Abundant specimens of this species are characteristic of the *semiforme* Zone in Hungary [167], although they are common also at lower levels [168]. The species is known as well from the middle Tithonian Substage of Cuba [169], Mexico ([75], Plate XXXII, figs. 3–11, Plate XXXIII), and Argentine [170]. According to Cecca *et al.*, [171], distribution of *A. neoburgense* is discrete: after residence in the middle Tithonian *"Burckhardticeras"* Zone, it appears again only in the Berriasian. Because of *Anaspidoceras* abundance in the *semiforme* Zone of West Europe, the *neoburgense* faunal horizon of the Russian plate corresponds, at least partly, to this stratigraphic unit.

13. *Pavlovia.* This Boreal species occurring from England and Greenland to the eastern slope of the sub-Polar Urals is common in the Volgian Stage of European Russia, but in Poland it is recorded only in some areas [172]. Therefore, *P. iatrensis* described by Zeiss from the upper Tithonian Substage of Austria ([46], p. 376, Plate 2, fig. 1) seems to be of a doubtful identification. As was mentioned, Mesezhnikov [50] discredited the original identification by Zeiss. Depending on its understanding, the *Pavlovia* genus proper can be characteristic of either the entire middle Volgian Substage, or its basal zone only. Therefore, if the generic classification by Zeiss is correct, one can conclude that the middle Volgian Substage corresponds partly to the upper Tithonian.

14. *Dorsoplanites.* Until recently, this genus was unknown from West Europe, being considered as characteristic mainly of the middle Volgian in Siberia, Greenland, and, to a lesser extent, in European Russia. One poorly preserved ammonite attributed to *Dorsoplanites* sp. has been figured by Khimshiashvili ([173], Plate 14, fig. 4) from the middle Tithonian Substage of Georgia. In addition, one of the species previously described by Schneid [57] from the middle Tithonian of the Neuburg locality was recently revised and attributed to this genus. Nevertheless, even if *Dorsoplanites lumbricatus* is correctly identified by Scherzinger and Schweigert [58], this species cannot be used for correlation. It is undoubtedly older than other known *Dorsoplanites* species, because it appears below first *Danubisphinctes* ex gr. *spurius* in the Neuburg section.

15. *Subdichotomoceras (Sphinctoceras).* Representatives of this subgenus are characteristic of the *subcrassum* Zone in the sub-Polar Urals and the Boreal realm in general. They are known from England (the *wheatleyensis* Zone), Greenland, and the Pechora River basin, being rare in southerly areas. In the Volga region, first *Subdichotomoceras* appear in the uppermost part of the *klimovi* Zone and are occasionally registered in the *sokolovi* Zone. The ammonites in question are mentioned also as present in the *hybonotum* (= *gigas* after Geyer [34]) Zone of southern Germany. In England, they appear in younger sediments implying their migration to Germany most likely from the Central Russian basin (via the Pripyat' strait). This event happened at the very beginning of the *sokolovi* Chron. The *Subdichotomoceras* occurrence in the basal Tithonian zone of Germany is consistent with data on distribution of *Paralingulaticeras* and with correlation between the lower part of the *sokolovi* Zone and the uppermost part of the *hybonotum* Zone.

16. *Franconites.* Mikhailov ([22], p. 56, Plate 11, fig. 1) described ammonites from the *sokolovi* Zone of central Russia under the name *Franconites* cf. *vimineus.* The ribbing of this species and changes in this character during ontogenesis are similar to those of typical *F. vimineus* described from the lower Tithonian *vimineus* Zone of southern Germany ([37], Plate 14). In this region, the genus *Franconites* proper characterizes the uppermost lower–basal middle Tithonian [174]. Recently, it was also found in the uppermost lower Tithonian Substage (the *darwini* Zone) of Spain [152], Hungary [43], Italy [175], and Mexico [176]. *Franconites* from the *sokolovi* Zone of the Russian plate, together with doubtful *Ilowaiskya pavida* and *?I.* cf. *pseudoscythica* from the *vimineus* Zone of the Franconian Alb, suggest that boundary between the *sokolovi* and *pseudoscythica* zones may be within the upper part of the lower Tithonian.

DISCUSSION AND PROPOSED CORRELATION SCHEME

At present, the following variant of correlation between the Volgian and Tithonian stages is widely usable: the lower Volgian Substage is correlated with the lower and middle Tithonian and the middle Volgian Substage with the upper Tithonian Substage [177]. Some researchers also correlate the *palmatus* and *ponti/"Burckhardticeras"* zones of the Tethyan areas [145]. It is clear, however, the last idea is inconsistent with data on distribution of *Lingulaticeras, Sutneria,* and *Pseudolissoceras.* The joint occurrence of these genera in the *panderi* Zone indicates that its lower part at least is correlative with the level not lower than the *fallauxi* Zone (Fig. 1).

Relationships between zones in different Tethyan areas are also controversial. None of the sections exhibits here the continuous *Danubisphinctes–Pseudovirgatites–Microcanthoceras* succession. The *Danubisphinctes–Pseudovirgatites* succession is known in Poland (Brzostówka section), and the

[17] I studied the original specimen from Semenov's collection stored at the Chair of Historical Geology of the St. Petersburg University.

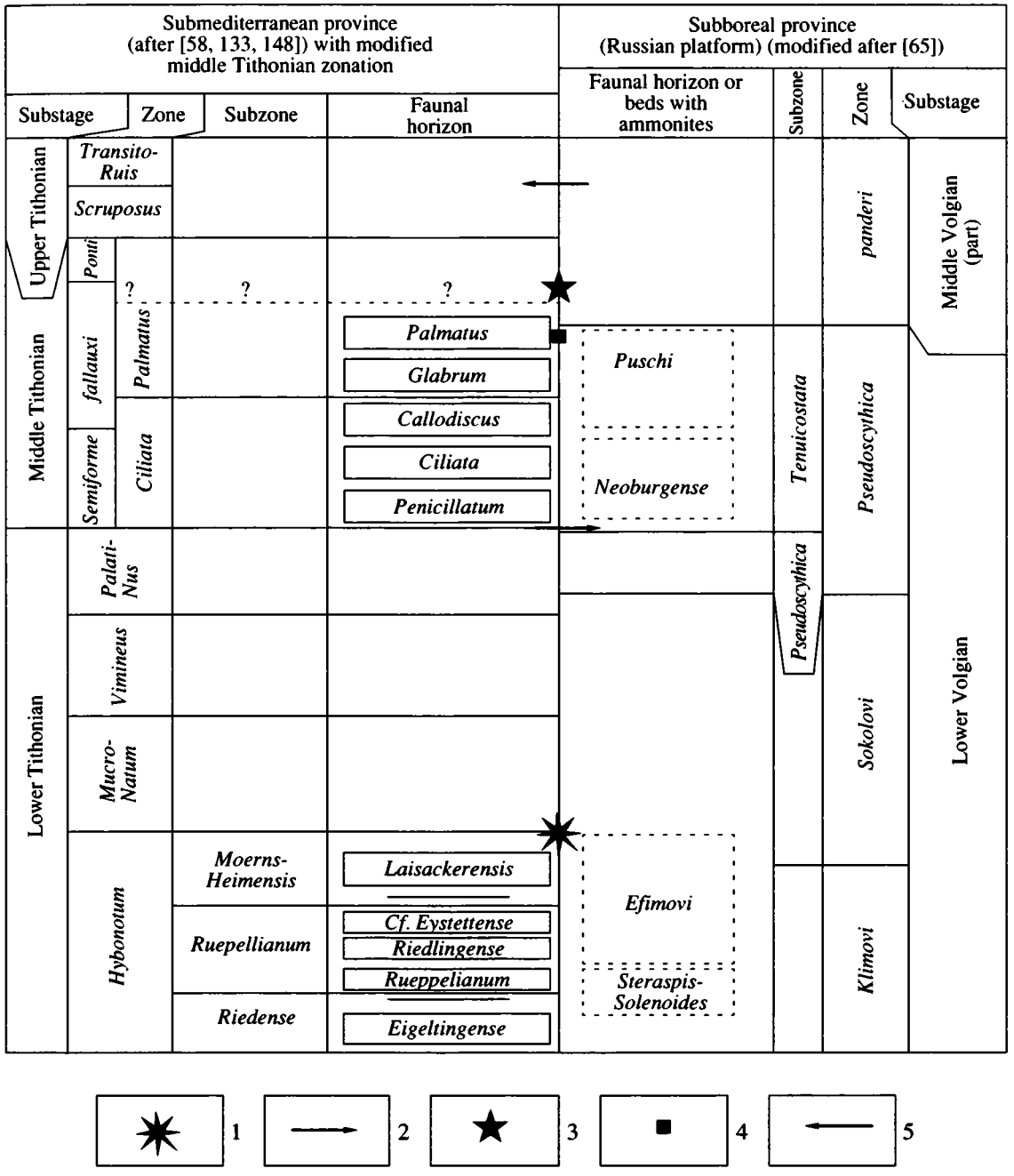

Fig. 1. Correlation of Volgian and Tithionian Stages based on ammonites with indicated changes in their assemblages most important for stratigraphic correlation: (1) disappearance of *Paralingulaticeras*, (2) appearance of *Anaspidoceras neoburgense* and *Sutneria asema* in the Russian plate, (3) disappearance of *Glochiceras* s. l., *Pseudolissoceras*, and *Sutneria*, (4) appearance of *Danubisphinctes* in Poland and Volga River sections, (5) appearance of *Zaraiskites regularis* in the Submediterranean Province.

Pseudovirgatites–Microcanthoceras succession is established in Spain, France, and the Carpathians. Moreover, the precise stratigraphic ranges are unknown for *Danubisphinctes* and *Pseudovirgatites* genera. Correlation of the *bavaricum* Zone with the entire middle Tithonian Substage [178] is also unjustified because of aforementioned data on distribution of *Pseudolissoceras* and *Sutneria*.

As it follows from the considered data on ammonite distribution, at least six levels directly correlative with the Tithonian zonation can be distinguished in the Volgian strata of European Russia (Fig. 1).

(1) The base of the *klimovi* Zone. Based on the appearance level of *Neochetoceras* cf. *steraspis* and on the disappearance level of *N. fridigense* (Berckh. et Hölder), the base of the *klimovi* Zone can be correlated with the base of the Tithonian *hybonotum* Zone.

(2) The base of the *efimovi* faunal horizon. The mass appearance of *Paralingulaticeras* forms replacing the

Neochetoceras–Lingulaticeras assemblage in the section corresponds most likely with similar changes in sections of southern Germany, where rare *Paralingulaticeras* appear together with first *Neochetoceras steraspis* in the *riedlingensis* faunal horizon [115]. The *efimovi* faunal horizon can probably be correlated with a part of the joint interval of cf. *eystettense* and *laisackerensis* faunal horizons of the Schwabian Alb and with Mörnsheimer Beds of the Franconian Alb.

(3) The lower part of the *sokolovi* Zone. *Paralingulaticeras* found in the *sokolovi* Zone of Poland and the Ul'yanovsk region near the Volga River suggest a partial correlation between this and *hybonotum* Zone, above which *Paralingulaticeras* is missing everywhere. The lower part of the *sokolovi* Zone likely corresponds completely or partially to the *laisackerensis* faunal horizon, the terminal one in the *hybonotum* Zone. This is consistent with *Subdichotomoceras* occurrence in the basal Tithonian of southern Germany.

(4) ?The upper part of the *sokolovi* Zone, which yields *Franconites* cf. *vimineus*, corresponds to that part of the synonymous zone in Germany, where *I. pavida* occurs (the last form characterizes mainly the upper part of the *sokolovi* Zone in Poland and Central Russia).

(5) The lower (?) part of the *pseudoscythica* Zone. *Ilowaiskya* cf. *pseudoscythica* occurs in Germany (Neuburg section) together with *Virgatosimoceras broilii*, the form characteristic of the *darwinni/albertinium* Zone. Therefore, the lower part of the *pseudoscythica* Zone corresponds most likely to the uppermost lower Tithonian.

(6) The *neoburgense* faunal horizon. Abundant *Anaspidoceras neoburgense* and *Sutneria asema*, both occurring in the middle part of the *pseudoscythica* Zone and in the middle Tithonian *semiforme* Zone, imply partial correlation between these stratigraphic units. It should be noted, however, that both species occur at lower levels of the Tithonian Stage as well.

(7) The *puschi* faunal horizon. The upper part of the *tenuicostata* Zone (or Subzone) of Poland yields *Danubisphinctes* characteristic of the uppermost ammonite-bearing level of the Neuburg Formation. Therefore, the *puschi* faunal horizon of the lower Volgian can be correlated, to a certain extent, with the *palmatus* faunal horizon of the Tithonian.

(8) The lower part of the *panderi* Zone. Basal layers of the *panderi* Zone contain *Lingulaticeras blaschkei, Pseudolissoceras*, and *Sutneria*. Inasmuch as all these forms never occur in Europe above the *fallauxi* Zone of the middle Tithonian, at least the lower part of the *panderi* Zone is not younger than the *fallauxi* Zone.

(9) The upper part of the *panderi* Zone. *Zaraiskites* ex gr. *regularis* from the upper Tithonian Substage of Poland and Bulgaria offers a possibility to correlate the upper part of the *panderi* Zone with a part of the *transitorius* Subzone of the upper Tithonian *microcanthus* Zone.

The scheme of correlation between the Volgian zones of Poland and the Tithonian zones of Neuburg Formation, which has been proposed by Scherzinger and Schweigert ([58], Fig. 1), can be accepted, though with reservations. In this scheme, the *tenuicostata* Zone corresponds approximately to the *palmatus* Zone (this is acceptable), but the *ciliata* Zone is shown to correspond to a hiatus in sections of Russia and Poland (this cannot be accepted). Views of these and French stratigraphers ([148], Table XIII) on relationships between different Tithonian zones are unacceptable. At least, the *ciliata* Zone, in which basal *penicillatum* and *ciliata* faunal horizons contain *Anaspidoceras neoburgense* (Oppel) and *Sutneria asema* (Oppel), cannot be younger than the *semiforme* Zone that also yields these species (Fig. 1). In turn, the *palmatus* Zone, which is correlated with the upper part of the *pseudoscythica* Zone of the Russian plate, is not equivalent to the *ponti* Zone [148] or to a part of it [58], and its upper boundary should be placed below that of the *fallauxi* Zone.

Until the correction by additional data, the upper boundary of the *panderi* Zone can be placed within the *transitorius* Subzone. In any case, the idea of Kutek and Zeiss [179] to correlate the last stratigraphic unit with the *regularis* faunal horizon only cannot be considered as well-substantiated, because the single (!) specimen of *Z. regularis* Kutek, 1994, found in the *transitorius* Subzone means nothing with respect to correlation of zonal boundaries.

Thus, the proposed variant of correlation between the lower Volgian and the Tithonian (Fig. 1) seems to be the most reasonable at present. I should note in addition that the reliable correlation with the Tithonian Stage is admissible at the moment only for the lower Volgian Substage coupled with the lower part of the *panderi* Zone. For the *virgatus–nodiger* interval, the direct Boreal–Tethyan correlation based on ammonites is impossible so far.

APPENDIX 1

DESCRIPTION OF *LINGULATICERAS BLASCHKEI* (CECCA ET ENAY, 1991) AND *SUTNERIA ASEMA* (OPPEL, 1865)

Many Mediterranean ammonites from the Volgian Stage of the Russian plate are poorly preserved and, unfortunately, they have never been described and figured. That is why I give below description of species *Lingulatriceras blaschkei* and *Sutneria asema*, which are important for the Boreal–Tethyan correlation. The described specimens are stored at the Paleontological Institute (PIN) RAS (collection no. 4861), the Vernadsky State Geological Museum (VSGM) (collection BX 17), and the Geological Institute (GIN) RAS (collections MIV and MK).

Suborder Haploceratina Besnosov et Michailova, 1983

Suprafamily Haplcerataceae Zittel, 1884

Family Haploceratidae Zittel, 1884

Subfamiliy Glochiceratinae Hyatt, 1900

Genus *Lingulaticeras* Ziegler, 1958

Lingulaticeras blaschkei (Cecca et Enay, 1991)

Plate, figs. 1–3

Oppelia strambergensis: Blaschke, 1911, [91] p. 154, Plate 1, fig. 7 (non fig. 6=*Neochetoceras strambergensis* Blaschke); Khudyaev, 1932, [181] p. 838, Plate 1, figs. 2, 3 (cf.); Khimshiashvili, 1957, [173] p. 55, Plate VII, fig. 3.

Streblitinae sp: Cecca *et al.*, 1990, [180] Plate 6, fig. 5.

"Glochiceras" blaschkei: Cecca and Enay, 1991, [128] p. 48, Plate 2, figs. 6–10, fig. 18 in text.

Glochiceras (?Lingulaticeras) blaschkei: Rogov and Egorov, 2002, [101] figs. 1, 2a, 2b.

non *Oppelia strambergensis*: Khudyaev, 1932, [181] p. 838, Plate IV, fig. 5 (=Streblitinae gen. ind.).

Holotype. Specimen FSL 162 510; figured in [128], Plate 2, fig. 7; France, Ardéche, La Pusin; middle Tithonian, *fallauxi* Zone, *richteri* Subzone.

Shape. Shell is discoid with a high oval cross section, narrow ventral side, and maximum thickness of the whorl in the lower third of the lateral side. The side furrow is noticeable only on the living chamber. Umbilicus is moderately narrow, step-wise.

No.	T	H	D	Du	T/H	T/D	H/D	Du/D
VSGM BX 17/1	4.8	10.3	20.4	4.5	46	23	50	22

Ornamentation is mainly represented by growth lines. Only in the upper part of the lateral side near the living chamber, there are rare, poorly distinguishable ribs. Some specimens have ribs slightly inclined forward, which appear already on phargmocone, in the lower part of its lateral side. Septal suture is slightly differentiated, though composed of relatively numerous elements (six lobes are distinguishable on the lateral side by the whorl height of 10 mm).

Comparison. In comparison with *?L. steueri* ([78], p. 20, Plate 1, figs. 3a, 3b), the species has larger dimensions and less distinct ornamentation. In distinction from *?L. umbilicocrenatum* ([98], Plate CXLIII, fig. 543), it is lacking developed costae in the lower part of the lateral side and has tapered ventral side. Some specimens of *L. blaschkei* ([128], Plate 2, fig. 6; this work, Plate, fig. 1) bear also well-developed internal robs, but they differ from *?L. umbilicocrenatum* by the arrow-shaped cross section of the test.

Distribution: the middle Tithonian (*fallauxi* Zone, *richteri* Subzone) of France, Italy, the North Caucasus, ?Crimea, Czechia (Stramberg) and the middle Volgian *panderi* Zone of the Russian plate (Chuvash Republic and Tver' oblast).

Notes. Ammonites from the Caucasus described as *Oppelia strambergensis* [182] are conventionally included into synonymy, since their ornamentation and lateral furrows cannot be observed because of a poor preservation. After examination of Khudyaev's original material stored at the TsNIGR Museum (collection no. 2925), I established that only one of figured specimens ([181], Plate 1, fig. 3; this article, fig. 1d) can be attributed, with sufficient confidence, to *L. blaschkei*. In distinction from it, other specimens, not figured inclusive, have rounded transverse section and relatively large dimensions of the test lacking the living chamber. In addition, septal sutures of these specimens are strongly differentiated, showing the large L lobe characteristic of the family Streblitinae. As was mentioned, *Lingulaticeras blaschkei* Cecca et Enay from France (Ardéche) is characteristic of a narrow stratigraphic interval within the *fallauxi* Zone that exhibits signs of the strengthened Tethyan influence. Migration of this species to the Central Russian basin was probably related to this episode.

Material. The collection includes the following well-preserved specimens: VSGM BX 17/1, middle Volgian Substage, *panderi* Zone, Yanyshar Ravine, outskirts of the Pervomaiskoe Village of the Batyrevskii area, Chuvash Republic, collection by V.V. Mitta; PIN no. 4861/25, middle Volgian Substage, *panderi* Zone, left bank of the Volga River 3 km downstream of the town of Kimry, Tver' oblast, collection by P.A. Gerasimov; middle Volgian Substage, *panderi* Zone (probably, lower part); TsNIGR Museum no. 40/2925, Tithonian Stage, North Caucasus, Tuapse area (original specimen by I.E. Khudyaev ([176], Plate 1. Fig. 3: *Oppelia strambergensis*).

Suborder Perisphinctina Besnosov et Michailova, 1983

Suprafamily Perisphinctaceae Steinmann, 1890

Family Aspidoceratidae Zittel, 1895

Genus *Sutneria* Zittel, 1884

Sutneria asema (Oppel, 1865)

Plate, figs. 5, 6

Ammonites asemus: Oppel, 1865, [183] p. 252.

Oppelia asema: Zittel, 1870, [184] p. 66, Plate 3, fig. 12.

Sutneria asema: Barthel, 1962, [150] p. 21, Plate 3, figs. 8–18; Hölder, 1964, fig. 73.10; Kutek and Wierzbowski, 1986, [66] p. 303, Plate 3, figs. 2, 3.

Sutneria (Sutneria) asema: Schlegelmilch, [185] 1994, p. 114, Plate 59, fig. 10.

Non *Eurynoticeras* aff. *asema*: Slavin, 1953, [149] p. 52, Plate 2, figs. 11–14 (=*Ptychophylloceras*) sp.

Holotype. Specimen AS III 54, figured in Zittel, 1870, [184] Plate 3, fig. 12; reproduced in Barthel, 1962, [150] Plate 3, figs. 8, 9; Schlegelmilch, 1994, [185] Plate 59, fig. 10; Poland, Rogoznik; Tithonian.

Shape. The shell is discoid, with oval cross section and rounded ventral side. The maximum thickness of

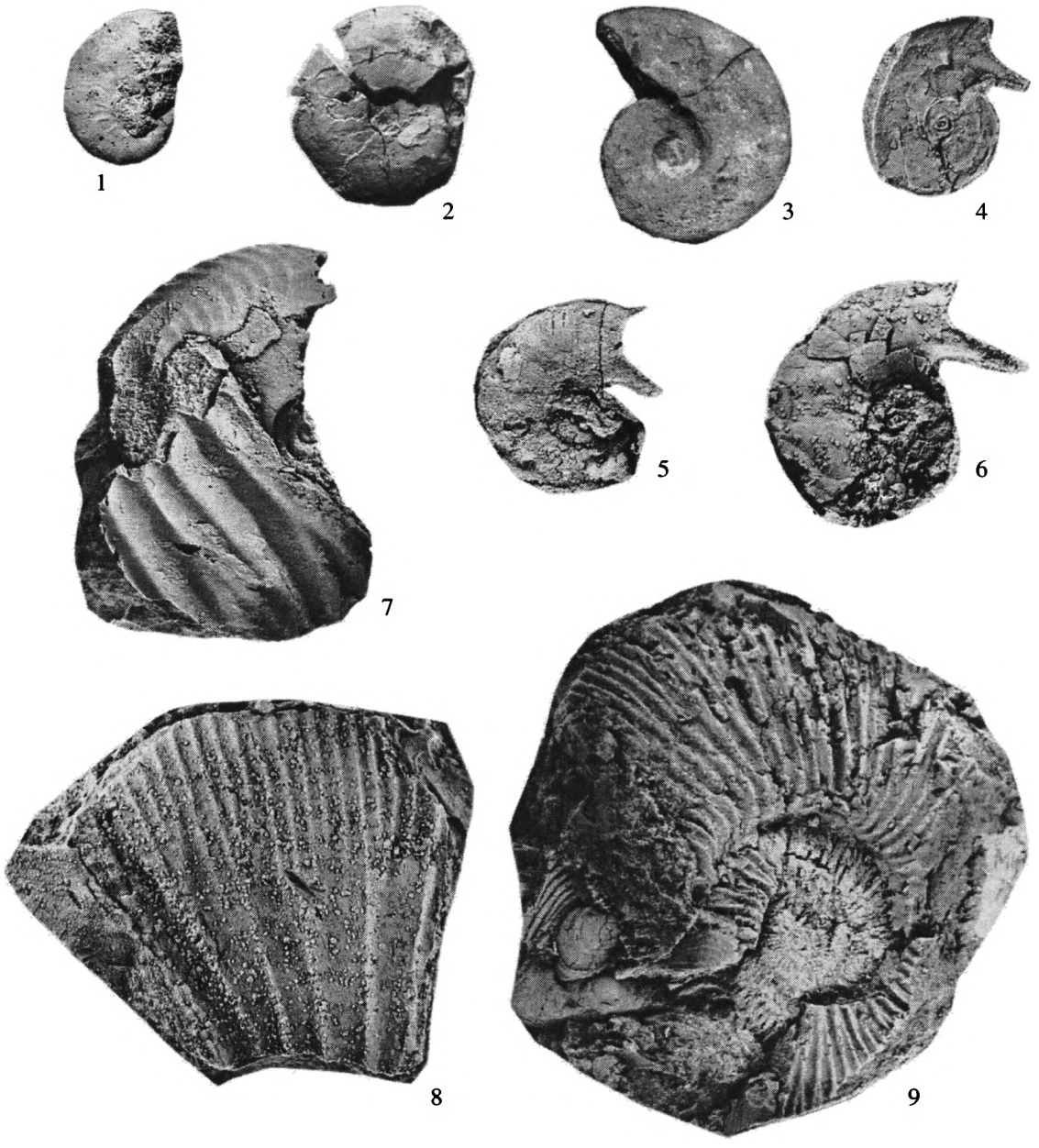

Plate. Some lower–middle Volgian ammonites important for stratigraphic correlation (real size, except for figs. 4–7 magnified ×2; specimens, if not specified otherwise, are from author's collection).

(1–3) *Lingulaticeras blaschkei* (Cecca et Enay, 1991): (1) Specimen PIN 4861/25; town of Kimry, Tver' oblast, collection of P.A. Gerasimov, (2) Specimen VSGM BX 17/1, Pervomaiskoe Village outskirts, Batyrevskii area of the Chuvash Republic, collection by V.V. Mitta, *panderi* Zone of the middle Volgian Substage, (3) Specimen TsNIGR 40/2925, original specimen figured in [181] (Plate 1, fig. 3: *Oppelia strambergensis*), North Caucasus, Tuapse area, Tithonian; (4) *Sutneria* cf. *eugyra* Barthel, 1959, Specimen MIV667/1, right bank of the Volga River near the Gorodishche Village, Ul'yanovsk area, lower Volgian Substage, *klimovi* Zone, Beds with *N. straspis–L. solenoids*, 0.7 m below the base of the *efimovi* faunal horizon; (5, 6) *Sutneria asema* (Oppel, 1865), right bank of the Volga River near the Gorodishche Village, Ul'yanovsk oblast, lower Volgian Substage, *pseudoscythica* Zone, *tenuicostata* Subzone, *neodurgense* faunal horizon: (5) Specimen MIV644, 0.48 m below the base of the *puschi* faunal horizon, (6) Specimen MK576, 0.45 m below the base of the *puschi* faunal horizon; (7) *Paralingulaticeras efimovi* (Rogov, 2002), quarry near the Murzitsy Village, Sechenovskii area of the Nizhni Novgorod oblast, lower Volgian Substage, *klimovi* Zone, *efimovi* faunal horizon; (8) *Pseudovirgatites* aff. *seorsus* (Oppel, 1865), Specimen MK594, right bank of the Volga River near the Gorodishche Village, Ul'yanovsk oblast, lower Volgian Substage, *pseudoscythica* Zone, *tenuicostata* Subzone, *puschi* faunal horizon, 0.6 m above the base of the *puschi* faunal horizon; (9) *Pseudovirgatites tenuicostatum* (Mikhailov, 1964), Specimen MK540, right bank of the Volga River near the Gorodishche Village, Ul'yanovsk oblast, lower Volgian Substage, *pseudoscythica* Zone, *tenuicostata* Subzone, *neoburgense* faunal horizon, 0.6 m below the base of the *puschi* faunal horizon.

whorls is in the lower third of the lateral side. Umbilicus is moderately wide, and umbilical wall is gently sloping. Aperture has well-developed elongated lappets and small near-umbilical constriction.

Ornamentation is differently developed in various specimens. Usually present are slightly developed retrocostate ribs in the upper part of the lateral side, which become more distinct on the ventral side and form a well-developed bench extending away from the aperture.

Comparison. In distinction from *S.* cf. *eugyra* (Plate, fig. 4), the described species has well-developed ribs only on the ventral side and is lacking ornamentation in the lower part of the whorl. Ornamentation patterns differ this species from older *Sutneria* forms.

Distribution: the lower (*darwini* and *vimineus* zones) and middle (*semiforme* and *ciliata* zones) Tithonian substages of France, southern Germany, Romania, Azerbaijan, and the Polish Carpathians; the Tithonian of Argentine; the lower Volgian (the *pseudoscythica* Zone, *tenuicostata* Subzone, *neoburgense* faunal horizon) of the Volga River basin. *Sutneria* forms from the lower part of the *panderi* Zone of the Gorodishche section probably belong to this species.

Notes. As is mentioned by consideration of *Sutneria* stratigraphic and geographic distribution, data on these ammonites from the Lower Cretaceous sediments [149] are doubtful. *Anaspidoceras neoburgense* used to be considered as antidimorph of *S. asema* is known however from the lower Berriasian. Therefore, it is probable that the stratigraphic range of *Sutneria* can be extended in future up to the Berriasian.

Material. Five well-preserved specimens (MK 560, 573, 576, 577, MIV 644) are from the Gorodishche section (Ul'yanovsk district of the Ul'yanovsk oblast; all samples are slightly deformed); specimen MK 609 is from a ravine located eastward of the Polevye Bikshiki Village (Batyrevskii area of the Chuvash Republic); lower Volgian Substage, *pseudoscythica* Zone, *tenuicostata* Subzone, *neoburgense* faunal horizon.

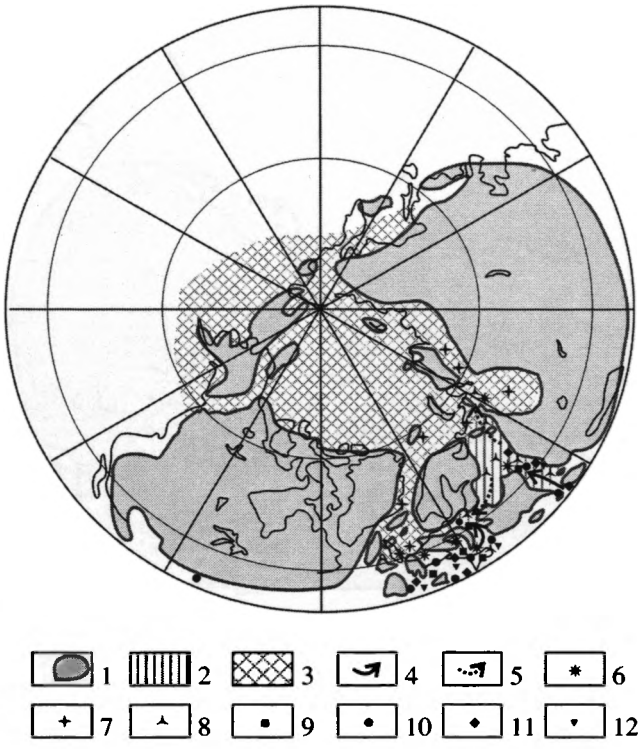

Fig. 2. Boreal–Tethyan migrations of ammonites during the *klimovi–sokolovi* chrons of the early Volgian (paleogeography after [198] modified): (1) land; (2) distribution area of Subboreal ammonites; (3) distribution area of Boreal ammonites; (4) northward and eastward migrations of Submediterranean ammonites; (5) migrations of Subboreal *Ilowaiskya* and *Gravesia* (*klimovi* Chron); (6) finds of *Gravesia* (*klimovi* Chron) and *Danubisphinctes* (*pseudoscythica–panderi* chrons); (7) finds of *Pectinatites*, *Subdichotomoceras* (early Volgian time), *Pavlovia*, and *Dorsoplanites* (middle Volgian time); (8) finds of *Ilowaiskya* (early Volgian time) and *Zaraiskites* (middle Volgian time); (9) finds of *Sutneria* and *Anaspidoceras* (*pseudoscythica* Chron); (10) finds of *Neochetoceras*; (11) finds of *Lingulaticeras*; (12) finds of *Haploceras* and *Pseudolissoceras*.

APPENDIX 2

BOREAL–TETHYAN AMMONOID MIGRATIONS IN THE NORTHERN HEMISPHERE DURING THE EARLY–MIDDLE VOLGIAN TIME

Many researchers who studied the lower–middle Volgian ammonites and stratigraphy discussed various aspects of paleobiogeography and possible migration paths of these organisms. The data generalized above specify distribution of ammonite taxa in the key areas of Volgian and Tithonian sediments and are important in this aspect.

During the early and initial middle Volgian time, the Central Russian sea basin was open, as repeatedly before, for ammonite migration from the Arctic, North Caucasian, and Polish basins [186]. Its unique position between the Panboreal and Tethys–Panthalassa superrealms is reflected in the peculiar composition of ammonites, which populated the Central Russian basin and have a high potential for the Volgian–Tithonian correlation. The detailed paleobiogeographic zoning of the basin is out of the scope of this work dedicated to mixed Boreal–Tethyan ammonite assemblages of the Northern hemisphere and to ammonite migration regardless of past biochores.

Based on distribution of ammonites, one can confidently suggest that the Central Russian and Polish sea basins communicated during the entire period under consideration, from the initial early to the middle Volgian time (at least to the *virgatus* Chron), although some researchers think differently [49].

Ammonite taxa whose migrations are discussed below have been either figured, or comprehensively considered above, as it was done, for instance, for *Pseudolissoceras* from the *panderi* Zone of the Volga

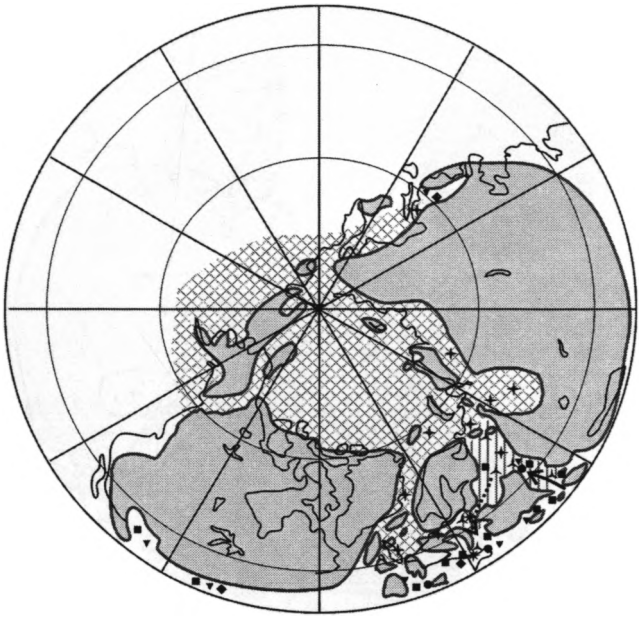

Fig. 3. Boreal–Tethyan migrations of ammonites diring the *pseudoscythica* Chron of the early Volgian (symbols as in Fig. 2).

River localities and for *Ilowaiskya* from the lower Tithonian of Hungary. In terms of paleogeographic interpretation, it is convenient to regard ammonite migrations coordinating them with the Subboreal ammonite scale of the Russian plate. Inasmuch as migrations of early Volgian ammonites were discussed recently in the other work [65], the main attention is focused this time on the middle Volgian migration events.

Since the very beginning of the Volgian Age, the role of Mediterranean and Subboreal ammonites was gradually decreasing in the Subboreal and Submediterranean seas, respectively. The fairly abundant Subboreal *Gravesia* and *Tolvericeras* occur in Germany only in the basal Tithonian (*hybonotum* Zone). Subsequent migrations of Subboreal ammonites to the Submediterranean province are evident only from single finds of their shells, which are frequently in a poor preservation state. *Gravesia* appeared in France, where the most complete *Tolviceras–Gravesia* succession is known [27], probably in the terminal Kimmeridgian. Representatives of this genus then migrated far to the east and penetrated in sea basins of Central Russia and sub-Polar Urals at the very end of the Kimmeridgian. Thus, these ammonites followed mainly the migration path similar to that of *Neochetoceras*. Although, *Gravesia* is unknown so far from Poland, the alternative migration path, for instance around Scandinavia, seems less probable. During the *klimovi* and *sokolovi* chrons, like in the Kimmerdgian, haploceratins penetrated in the Central Russian sea via the Pripyat strait, although some of them could migrate from the Caucasus (Fig. 2).

In addition to open seaways, one of the main factors responsible for ammonoid migration was probably an insignificant thermal gradient between waters of the Central Russian and Submediterranean seas. The mass migration of haploceratins to the Cantral Russian sea can also be explained by the fact that oppeliids dwelt in relatively deep areas of the basin, where the temperature gradient was lower than near the surface. However, *Paralingulaticeras* forms abundant in the *klimovi* Zone of the Russian plate were most likely shallow-water dwellers, because in southern Germany they occur in lagoonal sediments (Solenhofen and others localities).

Like in the terminal Kimmeridgian, when *Sarmatisphinctes* ([187], Plate 2, fig. 2) migrated from the Central Russian sea via Poland to Germany, Subboreal *Ilowaiskya* continued to migrate westward during the early Volgian. The last genus likely reached the Polish sea only, because its occurrence in Germany has been questioned recently [58]. Simultaneously, the abundance of *Ilowaiskya* decreased in the northeastern direction. In the sub-Polar Urals, only doubtful *Subplanites (Ilowaiskya)* sp. ([55], p. 86, Plate 4, figs. 2, 3), which may belong to *Pectinatites*, are known. *Ilowaiskya* from the Lena River lower course [188] have not been figured and are doubtful as well. Intense migrations of Boreal *Subdichotomoceras* southward and of Subboreal *Ilowaiskya* and *Gravesia* northward continued via the Timan–Pechora region at least during the *klimovi–sokolovi* chrons. Therefore, the assumption that the Central Russian sea had no connections with the Arctic basin [189] is inconsistent with the ammonite records.

The path of *Paralingulaticeras* and *Fontannesiella* migration during the early Volgian time in the Central Russian basin is unclear. Their absence in Poland, except for doubtful specimens in the *sokolovi* Zone, evidences against migration from the west. Both genera could probably migrate from the south, from the North Caucasian basin, but their absence in Jurassic sediments of the Orenburg oblast seems strange in this case. The dispersal of *Paralingulaticeras* and *Fontannesiella* in the Central Russian sea could probably be caused by a local warm current. No signs of Subboreal ammonites migration to the Caucasus during the early Volgian time are recorded.

In the greater part of the *sokolovi* Zone in Central Russia, molluscan assemblages are of a low diversity and consist of abundant *Ilowaiskya*, single *Subdichotomoceras* (ammonites), *Buchia*, and rare *Ostrea* forms similar to those from bivalve faunas of the sub-Polar Urals. The *sokolovi* Chron evidently marks the strengthened influence of the Arctic basin on the Central Russian sea. Simultaneously, the occurrence of rare Trigonidae and *Franconites* cf. *vimineus* in Central Russia is indicative of some warming episodes during this period.

Counter migrations of ammonites via the Pripyat' strait continued up to the end of the *sokolovi* Chron,

whereas later on, only migrations of Subboreal ammonites to Poland are registered [65]. In the platform part of Poland, southern ammonites (*Neochetoceras*) occur up to the top of the *sokolovi* Zone, replaced higher by prevalent Subboreal taxa. The sole exception is *Danubisphinctes* whose representatives penetrated into the Polish and then into the Central Russian seas at the end of the *pseudoscythica* Chron.

Ammonites abruptly changed their migration paths to the Central Russian sea during the *pseudoscythica* Chron (by the beginning of the *neoburgense* hemera). In addition to westward migration of *Ilowaiskya*, the mass invasion of Tethyan ammonites from the south happened owing to restoration of water exchange with the North Caucasian basin. Anyway, the complete absence of Mediterranean ammonites outside the Carpathian part of Poland suggests only this migration path that is consistent with occurrence of *Sutneria* and *Aspidoceras* s. l. in the Tithonian sediments of the Caucasus [190]. These thermophilic ammonites migrated to the Central Russian sea most probably with some warm currents, like during the *efimovi* hemera, because they are, for instance, more abundant and diverse in the Ul'yanovsk area near the Volga River and in the Chuvash Republic than in Jurassic sections of the Orenburg oblast located far away to the south (Fig. 3). On the other hand, during the *neoburgense* hemera, thermophilic elements from the Caucasian basin advanced farther to the north, up to the Moscow region, as compared with the earlier migration episode. It is noteworthy that substantial changes in the ammonite assemblages were accompanied by changes in other molluscan faunas: diversity of bivalves became almost an order of magnitude higher and belemnites turned out to be represented solely by thermophilic *Hibolites* (first in the Middle Volga region). Nevertheless, bottom waters remained probably relatively cold because *Buchia* were still abundant. This inference is consistent with data on benthic foraminifers, because their assemblages from the *pseudoscythica* Zone are similar to those from the *sokolovi* Zone [191].

During the *puschi* hemera, the character of ammonite migrations to the Central Russian sea slightly changed again. The influence of the North Caucasian basin became negligible, as it follows from changes in molluscan assemblages. Ammonites are represented only by rare *Pseudovirgatites* spp., which are known from the Polish basin as well, and by scarce *Danubisphinctes (?)*. Like at the beginning of the early Volgian, *Cylindroteuthis porrecta* appears among belemnites and *Buchia* forms amid bivalves. The strengthened Boreal influence is also recorded in West Europe, because *Pseudovirgatites tenuicostatum* appeared in Austria precisely at that time [46].

Northwest of the Polish sea, the early Volgian Subboreal population with *Ilowaiskya* and *Pseudovirgatites* was quickly replaced by the Boreal community with *Pectinatites* whose shells were found in ice-rafted boulders in Denmark ([192], *Perisphinctes (Zaraiskites)* cf. *scythicus*, p. 154, Plate 5, fig. 2; *P. (Z.) quenstedti*, p. 156, Plate V, fig. 3).

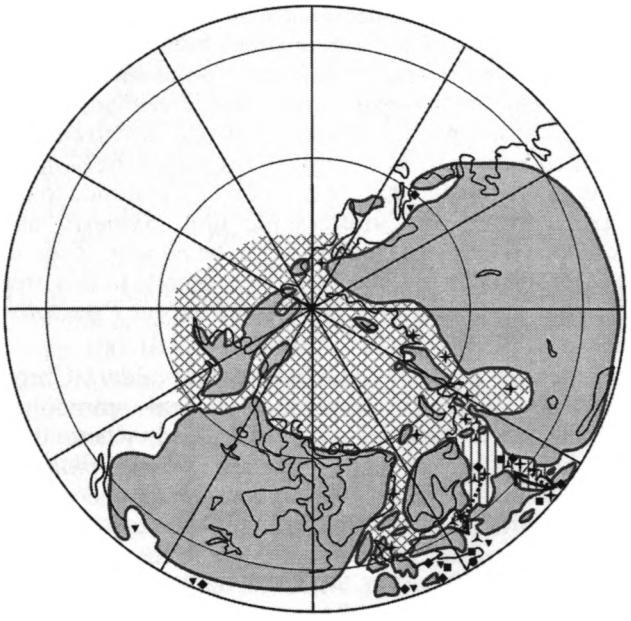

Fig. 4. Boreal–Tethyan migrations of ammonites diring the *panderi* Chron of the middle Volgian (symbols as in Fig. 2).

Typical Subboreal ammonites (*Ilowaiskya*) did not penetrate far northward: their single finds are known from the Pechora River basin and sub-Polar Urals region [193]. During the *pseudoscythica* Chron, Submediterranean ammonites were represented outside the Carpathian part of Poland only by *Danubisphinctes* species similar to those recorded in southern Germany. Several specimens of coarsely ornamented ammonites resembling *Danubisphinctes* were recently found in Poland, and this implies their penetration farther eastward.

Beginning from the middle Volgian time (*panderi* Chron), migration paths of ammonites changed again. Rare finds of Boreal *Dorsoplanites* are registered in the Caucasus, and some *Lingulaticeras blaschkei* migrated from this region as far northward as the 57°N (Fig. 4). This is indicative of a lower temperature gradient between the Central Russian and North Caucasian basins that was favorable for counter migration of ammonites. The Boreal–Tethyan migration of ammonites in this region was not significant, although a slightly strengthened influence of warm waters is again registered in the Orenburg oblast. The joint occurrence of *Ilowaiskya* and *Crassicolaria* is noted in the Mt. Khanskaya section [194]. If we assume that calpionellids are identified correctly, then we should admit the erroneous identification of ammonites, because such a high stratigraphic position of *Ilowaiskya* relative of the Tethyan zonal succession is inconsistent with all other available data. It seems that *Zaraiskites* forms were

taken for *Ilowaiskya* species. The lower part of the *panderi* Zone in the Orenburg sections bears signs of the strengthened Boreal influence: dominant among ammonites are *Dorsoplanites* and *Pavlovia*, while *Buchia* forms prevail among bivalves. *Zaraiskites* is abundant here only in the uppermost part of the *panderi* Zone. It is probable that *Crassicolaria* is confined to the upper part of this stratigraphic unit. *Sutneria* and *Pseudolissoceras* mentioned from the *panderi* Zone of the Gorodishche section could also migrate to this area from the south and, thus, one should expect *Pseudolissoceras* finds also in the Caucasus.

Like the early Volgian time, the *panderi* Chron marks a wide development of Boreal ammonites beyond the Carpathian areas of Poland, where they associate with endemic species *Danubisphinctes mazowiensis*. Some *Zaraiskites* forms migrated during the *regularis* hemera from this region southward to the Polish Carpathian and Bulgaria, where they coexisted with calpionellids (Fig. 4). Like the ancestral genus *Ilowaiskya*, *Zaraiskites* did not migrate from the Central Russian basin far northward. Between the Orenburg and Arctic regions, the latter were gradually replaced by Boreal *Dorsoplanites*, and only sporadic *Zaraiskites* finds are known in the Pechora River basin [195]. In this connection, the occurrence of *Dorsoplanites* instead of *Zaraiskites* in the Caucasus seems unexpected. *Haploceras* forms from the upper part of the *panderi* Zone of the North Caspian region, which have not been figured however [41], are the youngest Jurassic ammonites reliably evidencing the Tethyan influence on the Central Russian sea.

The last Jurassic episode of ammonites migration from the Central Russian basin was during the *virgatus* Chron. *Virgatites* is found in Poland [196] and *Lomonossovella lomonossovi* (not figured) is mentioned among ammonites from the Maikop area [197]. Penetration of thermophilic ammonites in Central Russia at that time is not recorded.

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New Data on the Marine Paleogene of the Southwestern Siberian Plate, Paper 1

M. A. Akhmet'ev*, G. N. Aleksandrova*, V. N. Beniamovskii*, D. I. Vitukhin*, Z. I. Glezer, Z. N. Gnibidenko***, V. D. Dergachev****, Zh. A. Dolya****, N. I. Zaporozhets*, G. E. Kozlova*****, I. A. Kul'kova***, I. A. Nikolaeva**, M. N. Ovechkina******, E. P. Radionova*, and N. I. Strel'nikova*********

**Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 119017 Russia*
***All-Russia Research Institute of Geology, St. Petersburg, Russia*
****Joint Institute of Geology, Geophysics, and Mineralogy, Siberian Division, Russian Academy of Sciences, Novosibirsk, Russia*
*****Omsk Expedition for Geological Prospecting, Omsk, Russia*
******All-Russia Research Institute of Petroleum Prospecting, St. Petersburg, Russia*
*******Moscow State University, Moscow, Russia*
********State University of St. Petersburg, St. Petersburg, Rùssia*

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Abstract—The Paleocene–Eocene marine deposits recovered by two reference boreholes (011BP and 9) in the Omsk depression, the southwestern Siberian plate, have been comprehensively studied in various paleontological aspects. Geology of the depression and associated Chany basin, which are structural elements of the Barabinsk lithologic–facies zone, is briefly described. Zonal assemblages of radiolarians, diatoms, ostracodes, benthic and planktonic foraminifers, organic-walled phytoplankton, spores and pollen are characterized. Biostratigraphic data are supplemented with lithological description of regional formations and with results of the electric logging and magnetostratigraphic study of Paleogene deposits that has been carried out for the first time in the West Siberian plate.

Key words: Paleogene, Eocene, Paleocene, West Siberian plate, foraminifers, dinocysts, radiolarians, diatoms, ostracodes, magnetostratigraphy, polarity zones, Gan'kino, Lyulinvor and Tavda formations.

INTRODUCTION

In addition to all the materials summarized in the correlation part of the unified regional stratigraphic scheme for Paleogene deposits of West Siberia [1], systematization of available paleontological data was used to substantiate the regional scales of the scheme. The results were published in a collective work by members of the Commission on the Paleogene System of the Interdepartmental Stratigraphic Committee [2] who emphasized that elucidation of time-and-space relations between the Gan'kino, Lyulinvor, and Tavda sequences and synonymous formations widespread in the study region is the key problem of the marine Paleogene stratigraphy. Unfortunately, the results of multidisciplinary study of boreholes 011BP and 9 drilled by the Omsk Expedition for Geological Prospecting in the south of West Siberia have not been timely considered in the full measure at the preparation time of the scheme and above publication. The paleontological study of both core sections has been accomplished recently, and the results obtained are presented in this paper. The members of scientific team who studied various groups of fauna and flora were G.E. Kozlova and D.I. Vitukhin (radiolarians); Z.I. Glezer, N.I. Strel'nikova, and E. P. Radionova (diatomaceous algae); I.A. Nikolaeva (ostracodes); G.N. Aleksandrova, N.I. Zaporozhets, and I.A. Kul'kova (organic-walled phytoplankton); V.N. Beniamovskii (foraminifers); and M.N. Ovechkina (nannoplankton). V.A. Krasheninnikov examined rock samples bearing planktonic foraminifers. Zh.A. Dolya and V.D. Dergachev were charged to select the borehole sites, to control the drilling and logging operations, and to compile and document the core sections. Z.N. Gnibidenko was responsible for the paleomagnetic research and magnetostratigraphic characterization of both reference sections drilled for the first time throughout the marine Paleogene of West Siberia and correlated with the time scale by Berggren *et al.*, [3]. M.A. Akhmet'ev and V.N. Beniamovskii headed the team and coordinated investigation in general.

A GEOLOGIC OUTLINE OF THE STUDY AREA AND RECOVERED CORE SECTIONS

Borehole 011BP has been drilled 1.5 km northeastward of the Achair Settlement, the Irtysh River bank

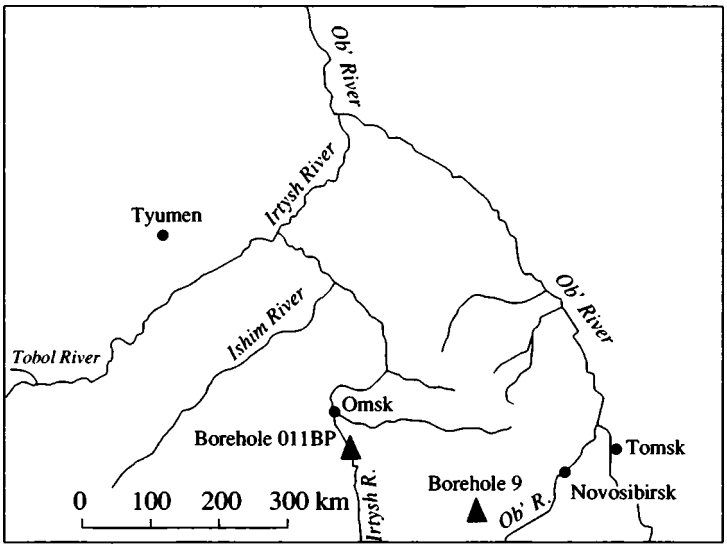


Fig. 1. Geographic localities of studied boreholes.

35 km upstream of Omsk, and Borehole 9 is situated near the Orlovka Village 25 km southeastward of Chistozernoe station (Omsk–Kulunda railway) at the southwestern coast of the Chany Lake (Fig. 1). The boreholes are drilled within the spacious Barabinsk lithologic-facies zone (or region) distinguished in the south of the West Siberian plate. In terms of tectonics, the greater zone part corresponds to the Omsk depression [4], and both boreholes are in the southern limb of the latter. The depression formation took place since the Mesozoic time. The Chany basin development within this large structure of the plate is recorded since the later Eocene time. The upper Lyulinvor Subformation and Tavda Formation are nearly as thick in the drilling area as in the most downwarped parts of the plate. The depth level of boundaries between the Paleogene lithostratigraphic subdivisions undulates depending on auxiliary local uplifts and downwarps (Chistozernoe depression, Kupa swell, etc.), development of which continued throughout the Paleogene [5]. Borehole 011BP is drilled near the Chany basin axis oriented parallel to the Irtysh River valley, where the crust downwarped most intensively during the Eocene. In the southern part of the Chistozernoe depression, where Borehole 9 is drilled, the depth to the Paleogene base ranges from 520 to 600 m. The situation around the Borehole 011BP is likely the same, although the Paleogene base has not been penetrated here. The depth to the Tavda Formation top is within the range of 250–280 m, and the total thickness of Paleogene marine deposits in the Chany basin is from 270 to 300 m.

The lithostratigraphy and biozonation of recovered sections are illustrated in Figs. 2–4. Drilling at the site 011BP was terminated at the depth of 498 m, and borehole penetrated here the complete Tavda Formation and the greater interval of the upper Lyulinvor Subformation. Borehole 9 reached the upper Maastrichtian clay of the Gan'kino Formation at the depth of 521 m. Oriented rock samples for paleomagnetic measurements have been taken with intervals of 0.5 to 1 m in between.

The recovered core sections are lacking rocks of the Talitsa Formation, and the Paleogene sequence begins at both sites, like in many other areas of the Barabinsk lithologic-facies zone, with opoka and sandstone beds of the lower Lyulinvor Subformation. The latter is overlain with a hiatus by the basal horizon of glauconite sandstone of the upper Lyulinvor Subformation. The horizon penetrated in the depth interval of 492.5–490 m is 2.5 m thick. In the core section 9 that is stratigraphically more complete, the upper subformation is divided in two members: the lower one composed of gray diatomaceous and opoka-like clay with rare opoka interlayers, and the upper member of greenish clay beds, which are less siliceous, bearing fish remains, and yield the diverse assemblages of diatoms, radiolarians, silicoflagellates, and dinocysts. Formerly, Shatskii [6] regarded the upper member as the individual Nyurol'ka Formation. In the core section 011BP, this member is almost completely eroded. Separate interlayers of greenish glauconite sandstones occur sporadically at different stratigraphic levels in both members. One of them most distinct in the core section 9 and confined to the boundary between the gray and greenish clay beds was penetrated in the depth interval of 451–450 m.

In the upper 2- to 3-m-thick interval of the upper Lyulinvor Subformation, clay is enriched in silty and sandy material prevailing over the clay matrix at the subformation top, especially in the core section 9. In the same section, the contact between Lyulinvor and Tavda formation is most distinct. At the base of the latter, there is a bed of glauconite sandstone and siltstone, which is 10 m thick and clearly depicted in the log, being correl-

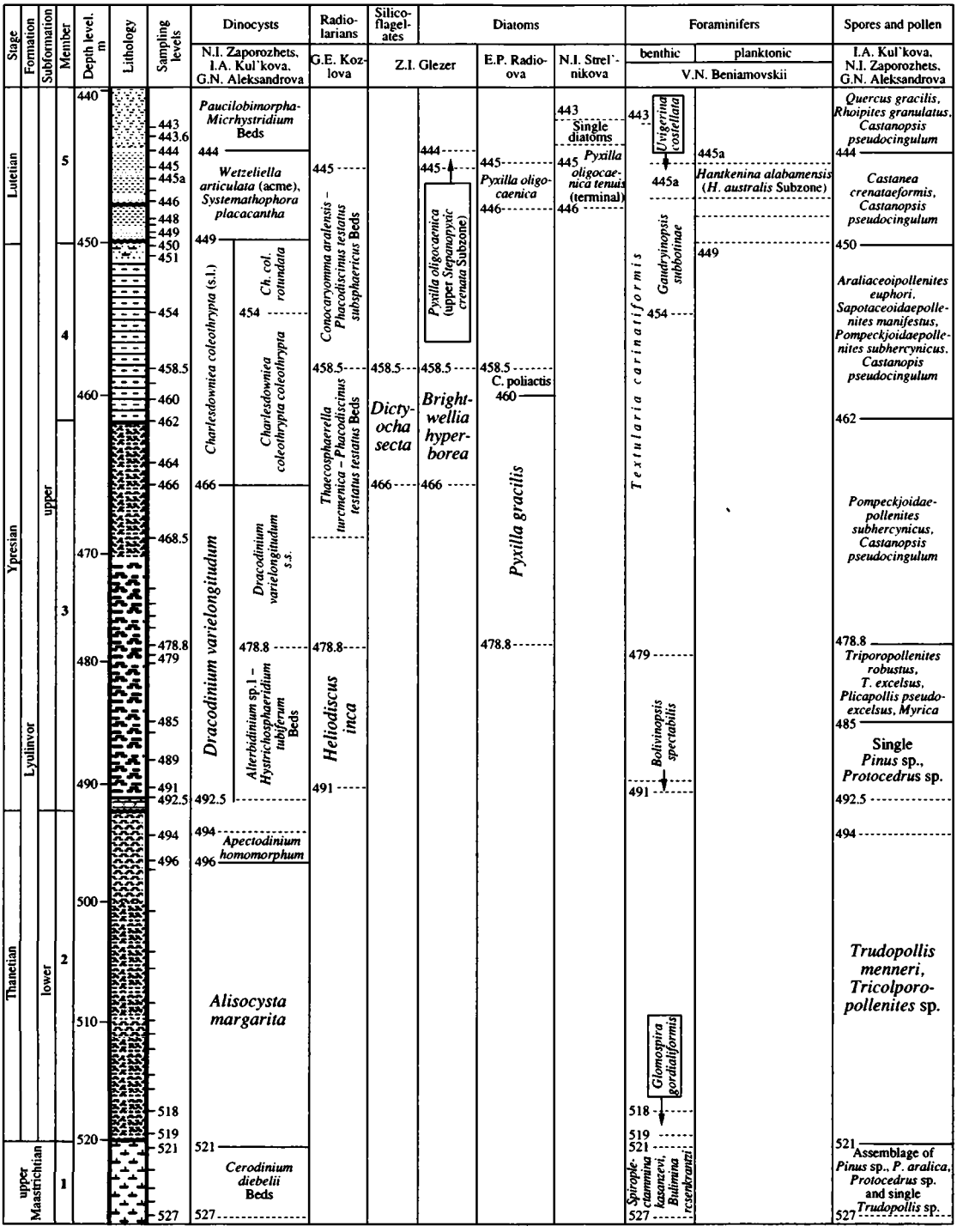


Fig. 2. Lithology and correlated biostratigraphic scales, Borehole 9, depth interval of 527–440 m (symbols as in Fig. 3).

ative with the regional logging marker "T." The rounded detritus of Lyulinvor clay occurs in the matrix of basal sands. In the core section 011BP, the contacts between formations and especially the Tavda basal member look differently. The latter is also 10 m thick but composed of greenish gray grumous aleurite with irregularly distributed clusters of clayey and silty particles. The member grades into overlying clay, and this transition looks gradual in the logging record as well.

The Tavda Formation is composed predominantly of clay with irregularly distributed admixture of silty material and with siderite interlayers and nodular seg-

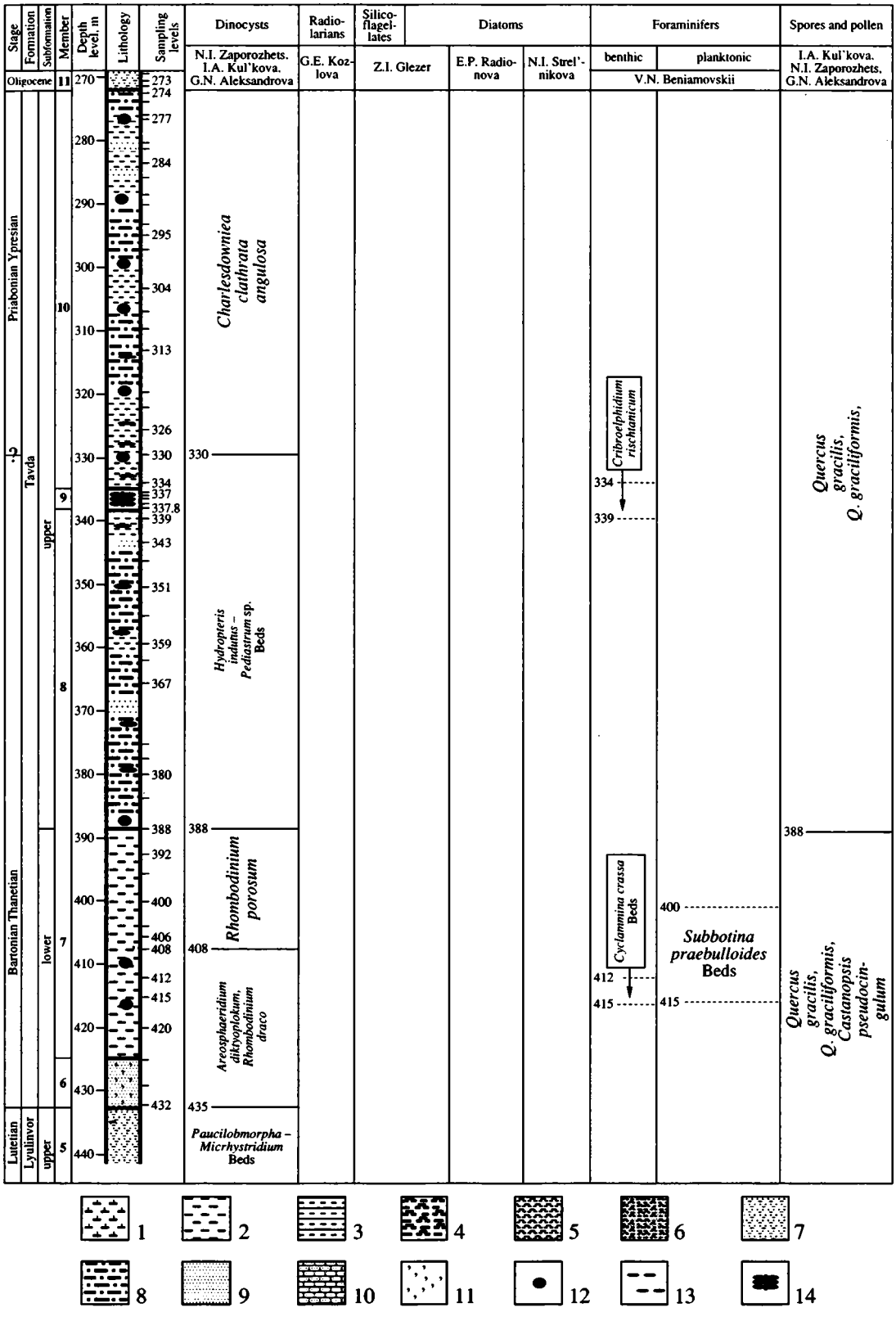


Fig. 3. Lithology and correlated biostratigraphic scales, Borehole 9, depth interval of 440–270 m: (1) calcareous, (2) non-calcareous, (3) siliceous, and (4) opoka-like clays; (5) opoka; (6) opoka-like sandy and (7) silty to sandy clays; (8) clayey siltstone or silty–sandy clay; (9) silty sand; (10) sandstone, siltstone; (11) glauconite; (12) septarian nodules of siderite; (13) marcasite; (14) leaf impressions of aquatic fern *Azolla vera* Krysht.

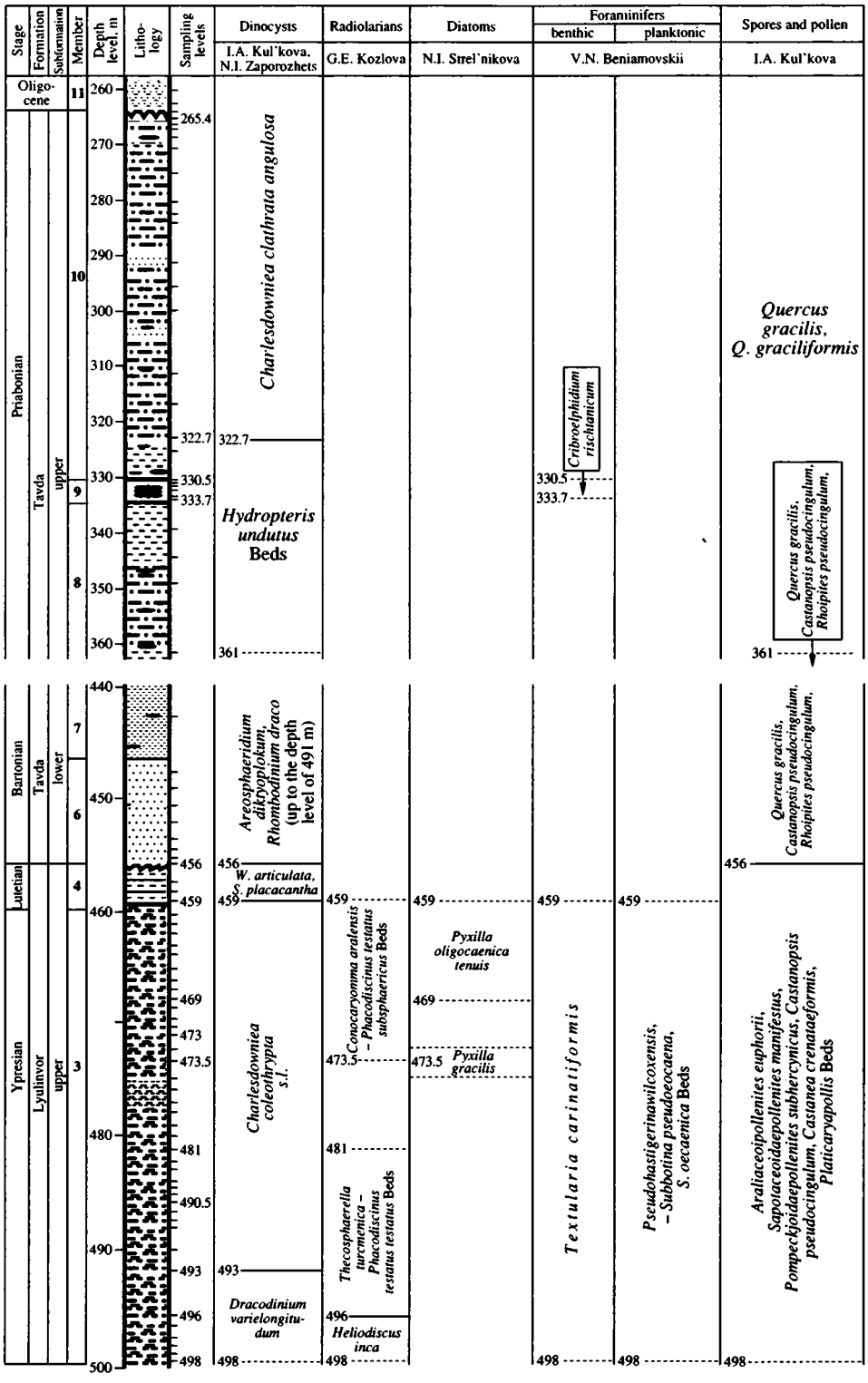


Fig. 4. Lithology and correlated biostratigraphic scales, Borehole 011BP, depth interval of 498–260 m (symbols as in Fig. 3).

regations. The purer clay is characteristic of lower and upper parts of the formation sequence, which characterize the true marine facies of the lower and upper Tavda subformations. Between these clay members, there is the middle one composed of silty clay. The three-member structure (if the basal sand bed penetrated by Borehole 9 is excluded) is readily distinguishable in both core sections. The Tavda Formation of the West Sibe-

rian plate used to be divided in two subformations, the upper of which includes the middle and upper members. We followed this tradition in our work as well. Nevertheless, the middle member composed of shallower facies yielding a specific assemblage of organic-walled phytoplankton can be regarded as an individual biostratigraphic subdivision. As it will be shown in the next our paper devoted to discussion of the results obtained, the core sections under consideration do not differ essentially from other sections recovered by nearby boreholes in the Barabinsk zone. All the sections are also very similar in parameters of electric logs.

PALEONTOLOGICAL AND MAGNETOSTRATIGRAPHIC CHARACTERIZATION

In order to compact presentation of materials, every lithostratigraphic subdivision, except for the Gan'kino Formation and lower Lyulinvor Subformation penetrated by Borehole 9 only, is characterized below based on data obtained for two core sections. Paleontological materials are uniformly presented beginning from data on the organic-walled phytoplankton that is of a high stratigraphic resolution and has been encountered in all studied samples.

Gan'kino Formation

The formation penetrated in the depth interval of 527–521 m is composed of massive to laminated clays bearing fragmented molluscan shells, echinoid spines, organic-walled phytoplankton, nannoplankton, foraminifers, and ostracodes.

The organic-walled phytoplankton (samples from depth levels of 525.5 and 522.5 m) is represented by 30 dinocyst species and by 8 species of prasinophytes and acritarchs (Table 1), which characterize the *Cerodinium diebelii* Beds of the mid-upper Maastrichtian. The dinocyst assemblage includes the following taxa: *Phanerodinium ?sonciniae, Ph. ?veligerum, Laciniadinium* cf. *firmum, Trythirodinium evittii, Kallosphaeridium* spp., *Membranosphaera maatsrichtica, Hystrichosphaeridium tubiferum, Microdinium ornatum, M. dentatum, M. kustanaicum, Chatangiella granulifera, Chatangiella* spp., *Fibradinium* sp., *Cerodinium speciosum, C. diebelii, Cerodinium* sp.1, *Isabelidium* sp., *Xenikoon australis, Connexrimura fimbriata, Pal eoperidinium pyrophorum, Exochosphaeridium* cf. *bifidum, Exochosphaeridium* sp., *Glyphanodinium facetum, Cordosphaeridium funiculatum, Parallecaniella* sp., *Achomosphaera ramulifera, Areoligera* sp.?, *Chlonoviella* sp., *Pterodinium cingulatum, Glaphyrocysta ordinata, Tanyosphaeridium* sp., *Alterbia?* sp., *Leberidocysta?* sp., *Cyclonephelium* sp.1, and *Hystrichostrogylon coninkii. Cerodinium* sp.1 present in the assemblage is close in morphology to *Deflandrea galeata*, being distinct from the latter by a large standard deltoid archeopyle characteristic of the genus *Cerodinium.* Acritarchs are represented by *Fromea chytra, F. laevigata, F. fragilis*, and *Micrhystridium* sp.

Nannoplankton is typical of the uppermost Maastrichtian zone CC26. About 50 species have been identified in rock samples from the depth levels of 527, 525.5, 524, 522.5 and 521 m. Most characteristic of them are *Lithraphidites quadratus, Micula mura*, and *Neophrolithus frequens.*

Foraminifers (the *Spivoplectammina kasanzevi–Bulimina rosenkrantzi* Zone of the upper Maastrichtian in West Siberia) represent an assemblage of benthic forms. The assemblage consists of the following species *Spiroplectammina kasanzevi, Gaudryina rugosa spinulosa, Dentalina cylindracea, Astacolus omskiensis, Globulina lacrima subsphaerica, G. veronica, Gyroidinoides obliquaseptatus, G. turgidus, Valvulinoides umovi, Eponides sibiricus, Cibicidoides bembix, Brotzenella pseudospinulosa, Cibicides gankinoensis, Anomalinoides pinguis, Bolivina plaita, Pseudouvigerina plummerae*, and *Bulimina thrihedra.*

Ostracodes of the Upper Cretaceous have been found in four studied samples (depth levels of 525.5, 524, 522, and 521 m). These are *Cetherella temporabilis, Cetherella* sp., *Pseudobythocythere verganica, Cythere (?) sibirica, Xestoleberis saratoensis, Krithe simplex, K. bronnemai, Bairdopillata simplicatilis, Curfsina notabilis, Clithrocytheridea schweyeri, Physocythere virginea*, and *Pterigocytheiris pallensa.*

Spores and pollen. Dominant in the palynological assemblage are *Pinus aralica* in association with other archaic species *Pinus* spp. and *Pseudocedrus* sp. Pollen *Trudopollis* sp. occurs as single grains.

Magnetostratigraphy. The normal and reversed polarities of rocks have been established respectively for the lower and upper parts of the clay member (apparent thickness 7 m) at the top of Gan'kino Formation (Fig. 5).

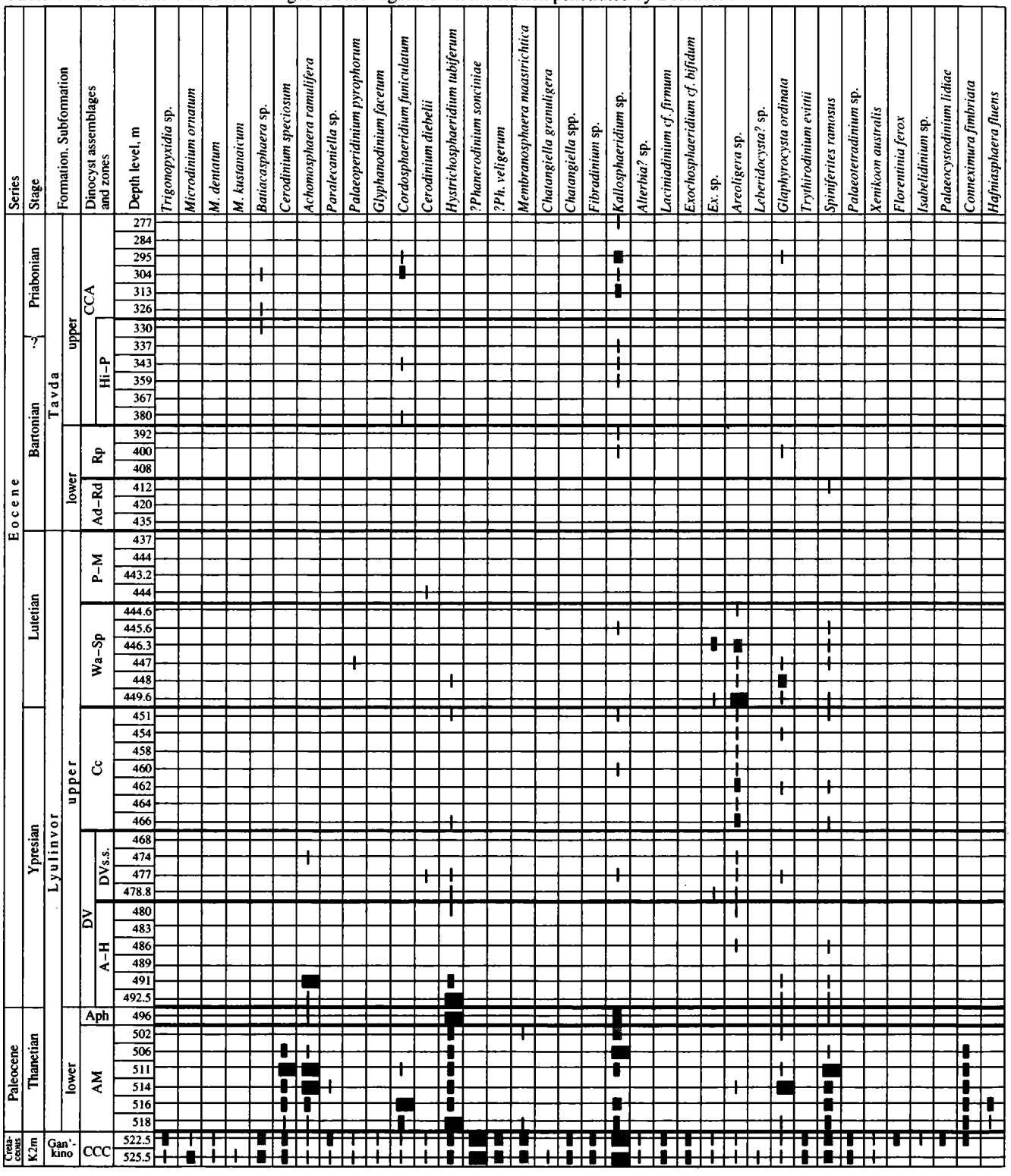
Electric and gamma-ray logging. In distinction from the Lyulinvor Formation, the more calcareous rocks of the Gan'kino Formation reveal a higher natural radioactivity of 8 versus 4 mr/h. The corresponding gamma-ray reading is less differentiated (Fig. 6).

Lyulinvor Formation

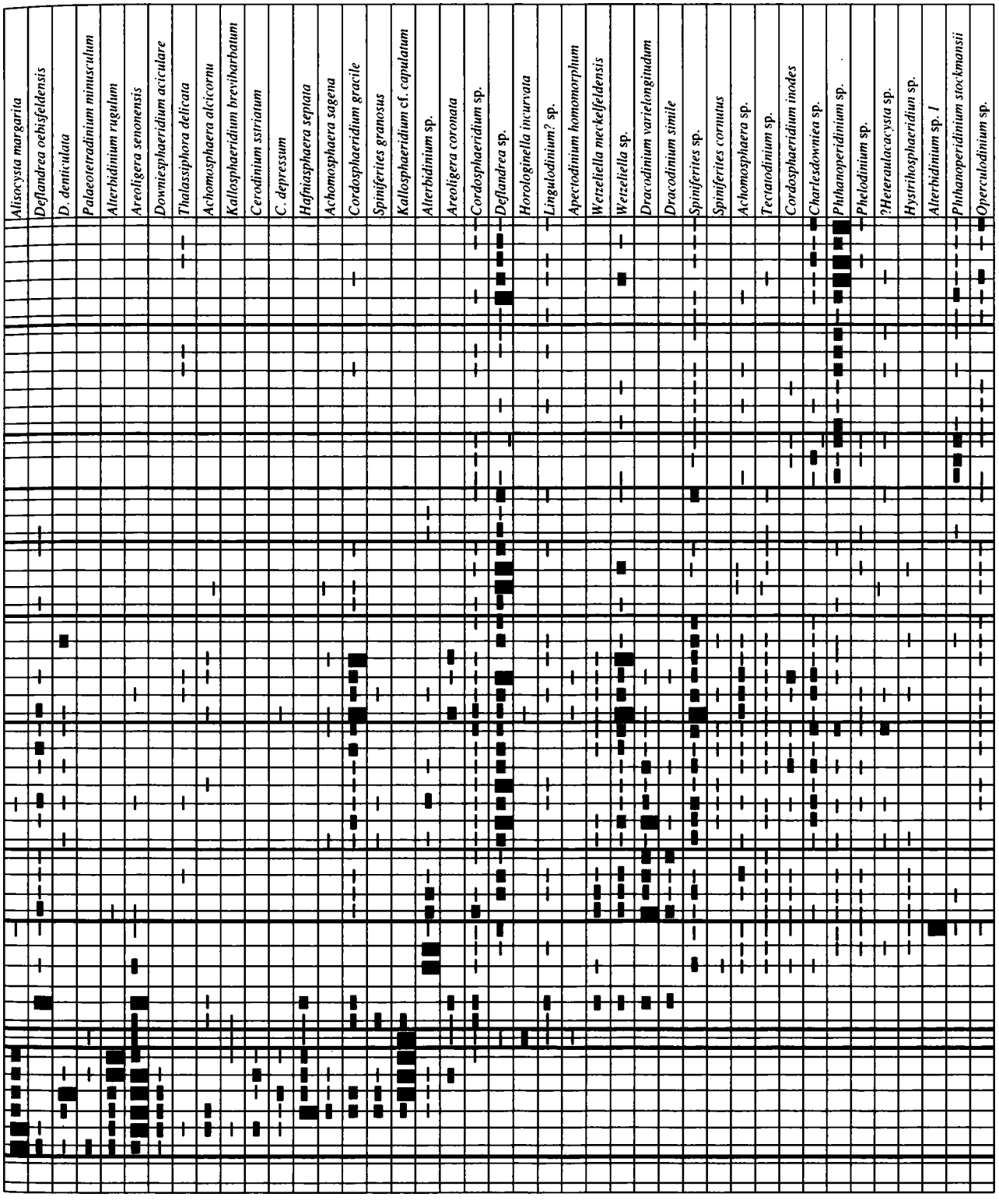
Lower Lyulinvor Subformation. Borehole 9 penetrated rocks of this subformation in the depth interval of 520.5–493 m. The unit is composed of alternating opoka, clay (opoka-like and diatomaceous), and sandstone beds. The basal bed of clayey bioturbated sandstone (1.5 m thick) rests on the erosion surface of the Gan'kino Formation. Organic remains are represented by foraminifers, organic-walled phytoplankton, spores, and pollen.

Organic-walled phytoplankton is macerated from 8 samples, the depth levels of 518, 516, 514, 511, 506, 502, 496, and 494 m. Over 40 dinocyst species and 8 prasinophyte and acritarch taxa are identified within the

Table 1. Zonation and distribution ranges of dinoflagellates in the section penetrated by Borehole 9



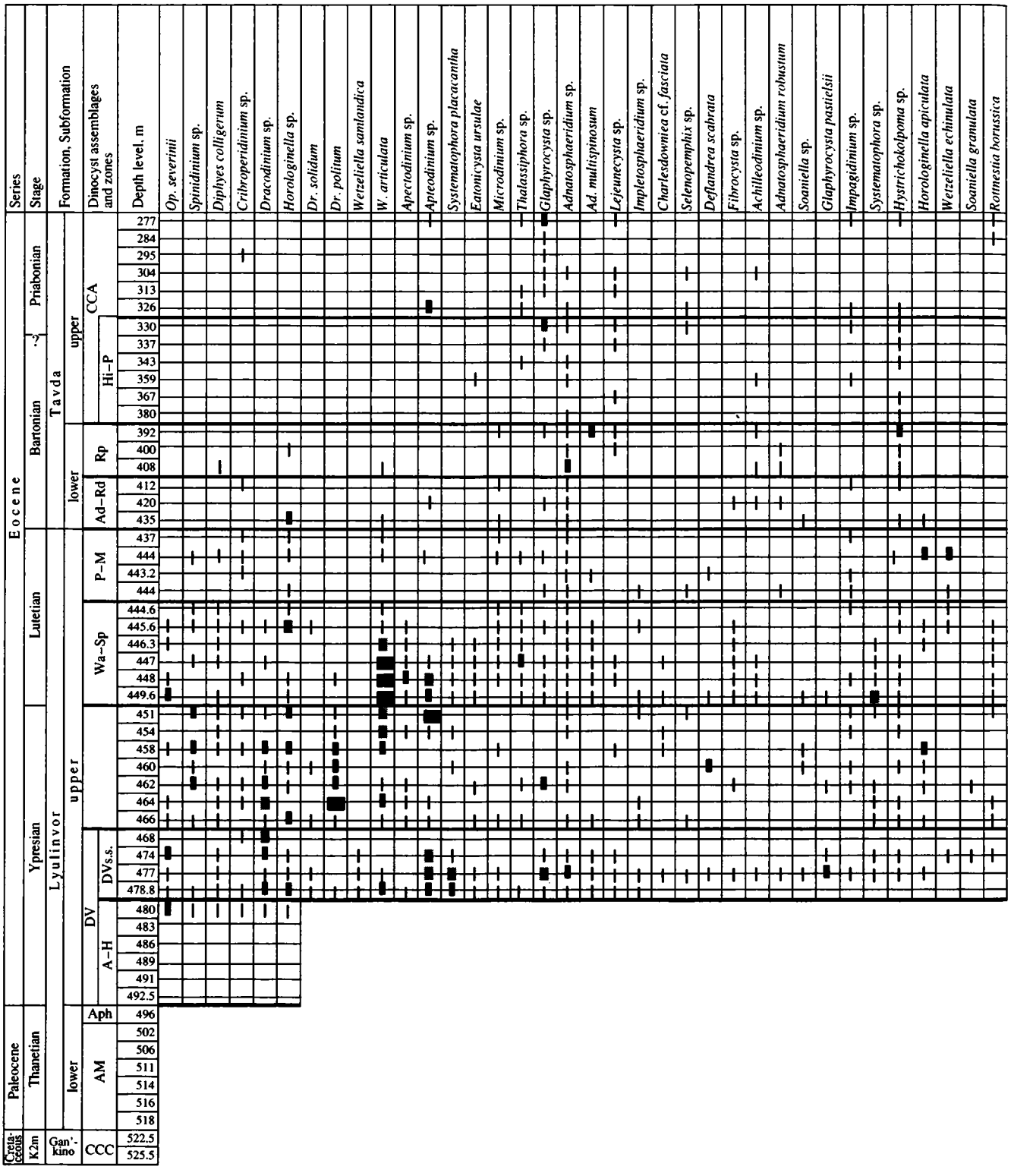
depth interval from 518 to 496 m (*A. margarita* Zone of the lower Thanetian). Distribution of phytoplankton forms is outlined in Table 1. Taxa dominating among dinocysts are *Alisocysta margarita, Alterbidinium rugulum, Cer(r)dinium speciosum, C. striatum, C. depressum, Deflandrea denticulata, D. oebisfeldensis, Areoligera senonensis, Areoligera coronata,* and species of *Cordosphaeridium, Spiniferites*, and *Achomosphaera* genera, while *Noremia* sp. and *Fromea* forms dominate among prasinophytes and acritarchs. *Apectodinium*



homomorphum and associated *Kallosphaeridium brevibarbatum, Kallosphaeridium* cf. *capulatum, Glaphyrocysta ordinata, Areoligera coronata, A. senonensis, Hystrichosphaeridium tubiferum*, and some other forms appear in the depth interval of 496–494 m (*A. homomorphum* Zone of the upper Thanetian). *Noremia* sp. is dominant taxon in the group of prasinophytes and acritarchs.

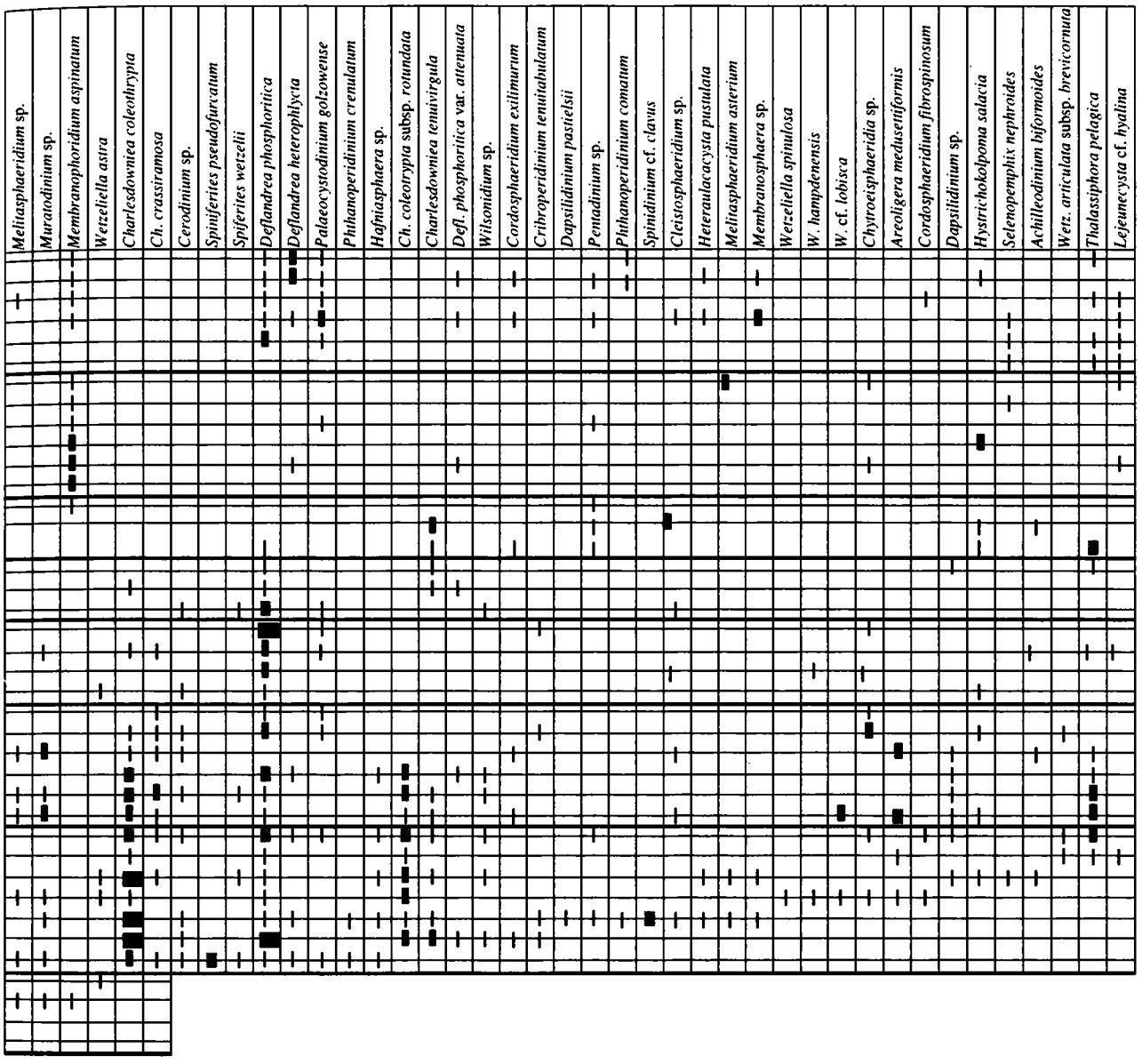
Foraminifers. Species *Glomospira* ex gr. *gaultina* and *G. gordialiformis* are identified in samples from the

Table 1. (Contd.)



depth levels of 519, 518, and 516 m (*G. gordialiformis* Beds). In the rank of zone, the beds were distinguished by Podobina [7] in the lower Lyulinvor Subhorizon of West Siberia.

Spores and pollen. Palynomorphs of higher plants (*Pinus* sp., *Protocedrus* sp., Taxodiaceae, *Trudopollis* sp., *Tricolporopollenites* sp., *Triporopollenites* sp., *Lygodium* sp.) occur as rare specimens.

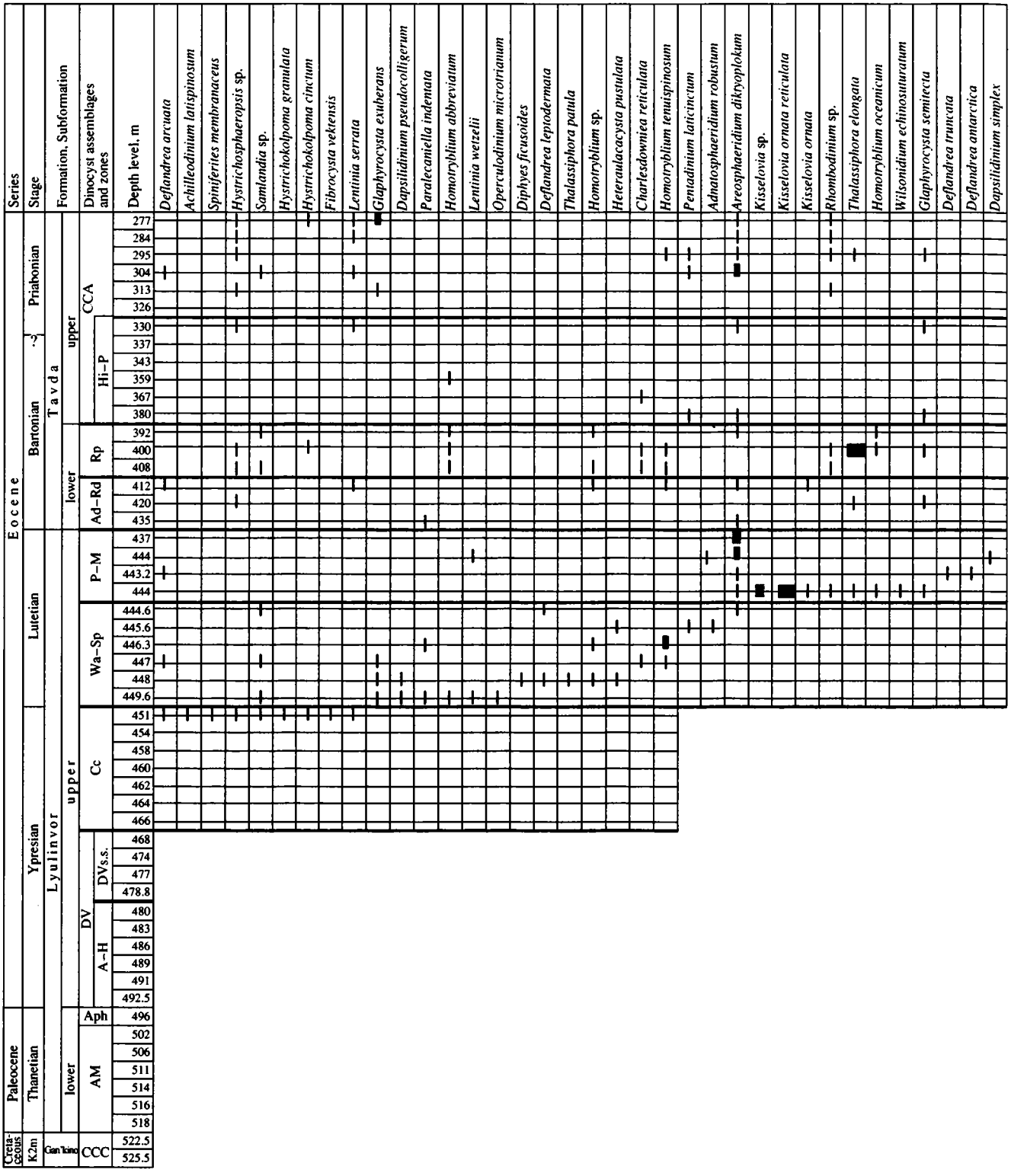


Magnetostratigraphy. The lower Lyulinvor Subformation spans four zones N_1ll, R_1ll, N_2ll, and R_2ll of normal and reversed polarity, which are 9, 5, 4, and 9 m thick, respectively. The lower (N_1ll) and upper (R_1ll) ones are longest in duration (Fig. 5).

Electric logging. Records in electric logs for the lower Lyulinvor Subformation are poorly differentiated. The apparent resistivity values (AR) correspond to 4–5 ohm. Spontaneous polarization (SP) is also low (3 to 4 mV). In the basal sandstone bed, the AR and SP parameters increase up to 12–15 ohm and 7–12 mV, respectively (Fig. 6).

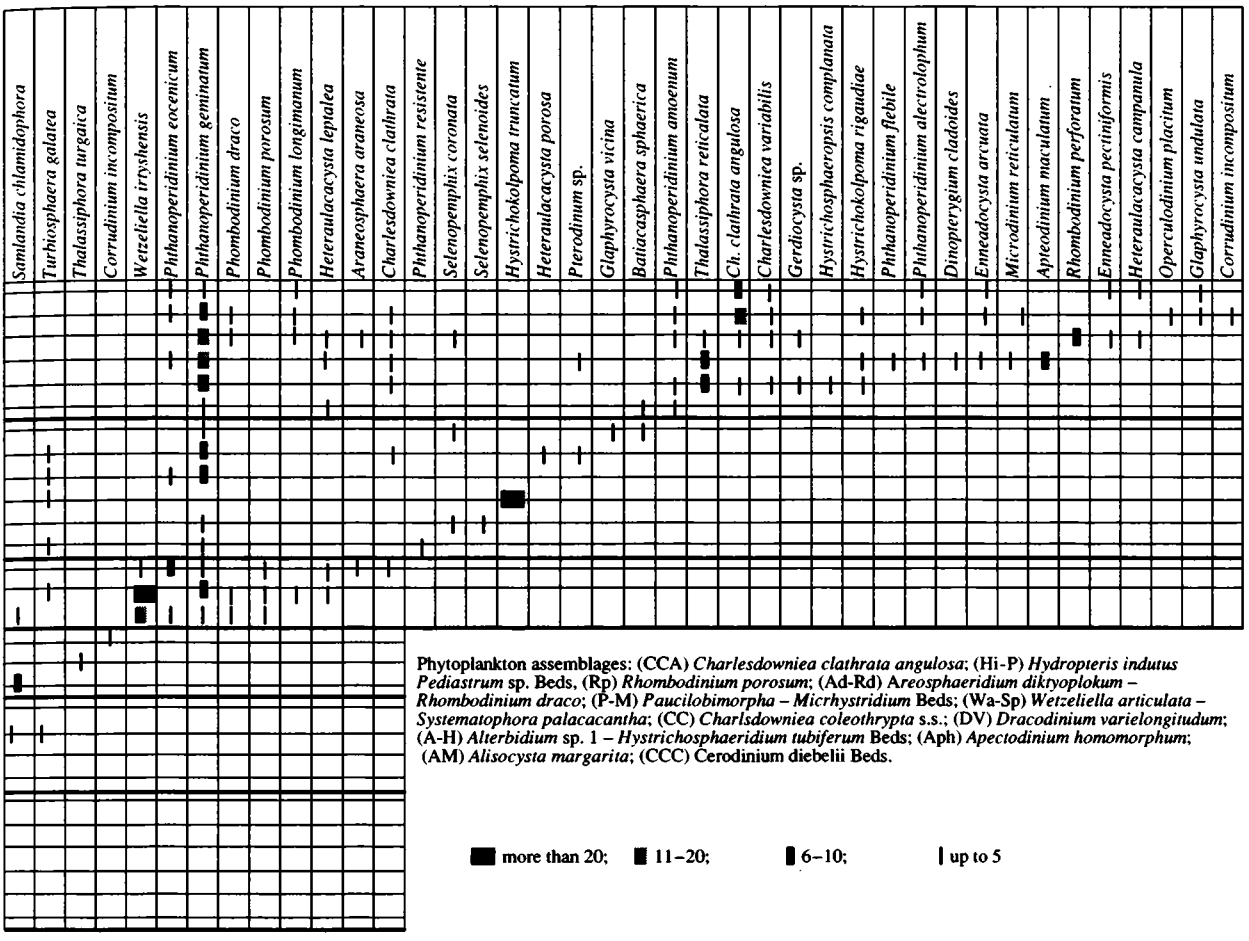
Upper Lyulinvor Subformation. The subformation penetrated by Borehole 9 in the depth interval of 493–435 m is 58 m thick. Its basal bed of fine-grained glauconite sandstone overlies the erosion surface of underlying deposits. Distinguishable in the rest of the sequence are two members. The lower Member "a" of light gray interlayering diatomaceous and opoka-like clays (depth interval of 493–451 m) encloses a bed of sandstone with cherty matrix, which is 1 m thick (depth interval of 470–469 m). The overlying Member "b" is composed of dark green clays (depth interval of 450–435 m), which bear scales and fine fish detritus and are intercalated with sandstone and siltstone interlayers up to 1 m thick (depth intervals of 451–450 and 448–447.5 m). Near the top, the clays are enriched in sandy and silty material. As is shown below, the upper half of Member "b" can be distinguished based on organic-walled phytoplankton as a separate biostratigraphic unit that is lacking, however, the distinctive lithologic features. In the core section 9, the subformation rocks yield the

Table 1. (Contd.)



diverse assemblages of various microplankton groups (diatomaceous algae, radiolarians, benthic and planktonic foraminifers, organic-walled phytoplankton, spores and pollen).

Borehole 011BP penetrated only an upper portion of Member "a" with the hardground surface at the top. Drilling at this site was terminated at the depth of 498 m. The top of the upper Lyulinvor Subformation was



Phytoplankton assemblages: (CCA) *Charlesdowniea clathrata angulosa*; (Hi-P) *Hydropteris indutus Pediastrum* sp. Beds, (Rp) *Rhombodinium porosum*; (Ad-Rd) *Areosphaeridium diktyoplokum – Rhombodinium draco*; (P-M) *Paucilobimorpha – Micrhystridium* Beds; (Wa-Sp) *Wetzeliella articulata – Systematophora palacacantha*; (CC) *Charlsdowniea coleothrypta* s.s.; (DV) *Dracodinium varielongitudum*; (A-H) *Alterbidium* sp. 1 – *Hystrichosphaeridium tubiferum* Beds; (Aph) *Apectodinium homomorphum*; (AM) *Alisocysta margarita*; (CCC) Cerodinium diebelii Beds.

penetrated here at the depth of 456.2 m, and the recovered interval of this subdivision is as thick as 41.8 m. Like in the core section 9, Member "a" is composed of interlayering opoka-like and silty clays, which are light gray in the upper part and colored darker near the base of the recovered interval. Interlayers of fine-grained sandstone and siltstone (0.1 to 0.2 m thick) have been penetrated at the depth levels of 486.4 and 471 m. Like at the other site, the content of silty material, dispersed glauconite grains included, increases in the upper 4-m-thick interval of the sequence close to the contact with the overlying Tavda Formation. This interval correlative with Member "b" of the core section 9 is partially preserved after erosion. Rocks of the upper Lyulinvor Subformation, which have been recovered by Borehole 011BP, also yield abundant microplankton.

Organic-walled phytoplankton. Aleksandrova and Zaporozhets studied 26 samples: 15 from Member "a" and 11 from Member "b" of the core section 9 sampled at the depth levels of 492.5, 490, 489, 486, 480, 478.8, 477, 474, 468, 466, 464, 462, 460, 458, 454, 451, 449.6, 448, 447, 446.3, 445.6, 444.6, 444, 443.2, 440, and 437 m. In addition, Kul'kova studied 10 samples from the depth range of 490–440 m, where they were taken with the in-between intervals of 5 m. Dinocysts (10–12 to 70–75 species) were macerated from all the samples, some of which also yielded abundant prasinophytes and acritarchs. The distinguished dinocyst assemblages characterize the *Dracodinium varielongitudum* s. l. Zone of the middle Ypresian (492.5– 466 m), the *Charlesdowniea coleothrypta* s. l. Zone of the upper Ypresian–lower Lutetian (466–451 m), and the *Wetzeliella articulata* (acme)–*Systematophora placacantha* Zone of the upper Lutetian (449.6–444 m). The former two zones are distinguished in Member "a," and the third one spans the lower half of Member "b." The interval between depth levels 444 and 435.5 m is ranked as the *Paucilobimorpha* (*Micrhystridium*) Beds according to their dominant taxa representing up to 70% of the organic-walled phytoplankton assemblage.

In the core section 011BP, Kul'kova studied 15 samples (depth interval of 497.7–454.8 m) from the upper Lyulinvor Subformation, and Zaporozhets additionally investigated 5 samples from the depth levels of 498, 490.5, 470.8, 459, and 456 m. They distinguished the uppermost part of *Dracodinium varielongitudum* Zone (samples from depth levels of 498 and 493.7 m) and the overlying *Charlesdowniea coleothrypta* Zone (493.2–

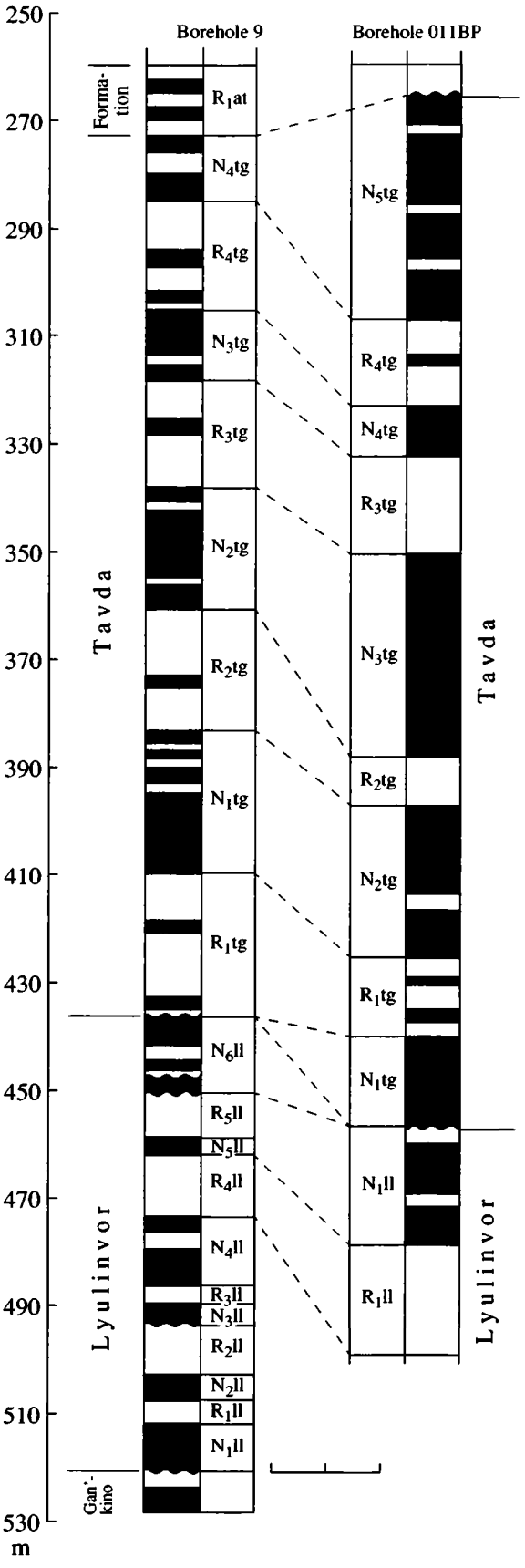


Fig. 5. Magnetostratigraphy of boreholes 9 and 011BP.

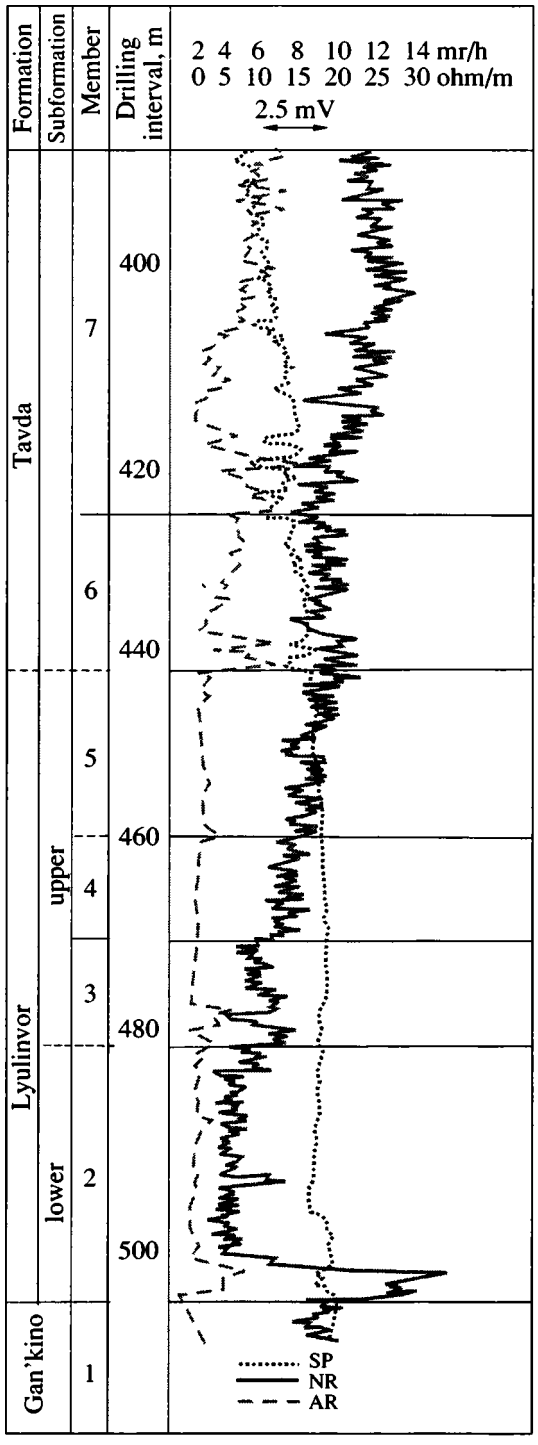


Fig. 6. Electric logs, Borehole 9, depth interval of 510–390 m: (SP) spontaneous polarization; (NR) natural radioactivity; (AR) apparent resistivity.

?460 m). In opinion of Zaporozhets, two samples from depth levels of 459 and 456 m belong to the *Wetzeliella articulata* (acme)–*Systematophora placacantha* assemblage, while Kul'kova believes that they still exemplify the preceding *Charlesdowniea coleothrypta* s. l. Zone.

In the core section 9, the *Dracodinium varielongitudum* Zone spans the depth interval of 492.5–466 m. Its base corresponds to the erosion surface between the lower and upper Lyulinvor subformations. The first occurrence level of *Dracodinium varielongitudum* (15 grains) is established at the depth of 491m. The zone is divided in two parts (subzones?) termed as the *Alterbidinium* sp.1–*Hystrichosphaeridium tubiferum* (492.5–478.8 m) and *Dracodinium varielongitudum* s. str. beds (478.8–466 m). In the lower beds, their index species is rare, whereas other indicative forms of the zone represent up to 50% of the spectrum in some samples. The mass abundance of the index species is first recorded at the depth of 478.8 m only, and the lower part of this biostratigraphic subdivision can be regarded as transitional from the lower *Wetzeliella meckelfeldensis* and *D. simile* zones, which are missing from the studied succession however. It is also remarkable that *Alterbidinium* sp. and *H. tubiferum* disappear almost completely in the interval of *D. varielongitudum* predominance. The lower zone boundary (the first occurrence level of index species) is recorded at the depth of 491 m above the hiatus in the section. Nevertheless, it is logical to attribute to this zone the interval of sands (sample from the depth of 492.5 m) occurring below this level and down to the erosion surface. The appearance level of frequent *Charlesdowniea coleothrypta* defines the upper boundary of the zone.

In addition to index species, typical of the zone under consideration are *Glaphyrocysta ordinata, G. pastielsii, Hafniasphaera septata, Deflandrea oebisfeldensis, Hystrichosphaeridium tubiferum, Achomosphaera crassipellis, Cordosphaeridium biarmatum, Horologinella inculta*, and some other forms. Species *Spiniferites granomembranaceus, Cerodinium depressum*, and *Operculodinium uncinispinosum* are characteristic of this zone only and do not occur above it. Taxa characteristic of older assemblages occur in the lower interval and may be partially redeposited. These are either *Areoligera senonensis, Alterbidinium* sp.1, *Kallosphaeridium brevibarbatum* and other *Kallosphaeridium* forms, or species limited in distribution by the upper zonal boundary and occurring as single specimens above it. The *Cordosphaeridium, Achomosphaera, Spiniferites*, and *Deflandrea* species mostly belong to transit taxa. *Noremia* and *Cometodinium* forms are characteristic representatives of prasinophytes and acritarchs from the lower part of the zone. Beginning from the depth of 478.5 m, there is recorded a mass abundance of *Dracodinium* species (*D. varielongitudum, D. condylos, D. politum*). Other encountered forms are *Wetzeliella articulata, Charlesdowniea edwardsii, Eatonecysta ursulae, Adnatosphaeridium multispinosum, Apteodinium* spp., *Rottnestia borrusica, Soanella granulata, Thalassiphora delicata*, and single specimens of *Tectatodinium* sp., *Lingulodinium* sp., *Impagidinium* sp., *Spinidinium* sp., and *Selenopemphix* sp. The dinocyst assemblage from *Dracodinium varielongitudum* Beds of the core section 011BP is very similar in composition to that from the core section 9.

The lower boundary of the *Charlesdowniea coleothrypta* Zone (Borehole 9, depth interval of 466–451 m; Borehole 011BP, depth interval of 493–459 m) is placed at the first occurrence level of its index species, and the upper one is established at the depth of 451 m, where the index species is still rather frequent (especially subspecies *Ch. coleothrypta rotundata*). Nevertheless, noticeable changes in composition of dinocysts are recorded at this level, where a sandstone interlayer (depth interval of 451–450 m) has been deposited in response to a sudden change in sedimentation environment. Within the *Charlesdowniea coleothrypta* s. l. Zone, distribution of *Ch. coleothrypta* s. s. and *Ch. coleothrypta rotundata*, the index taxa of synonymous subzones, is established conventionally. *Ch. coleothrypta rotundata* appears 10 m above the zone base, and its abundance rate is much lower than that of *Ch. coleothrypta* up to the depth of 454 m. In the upper beds of zonal sequence (samples from depth levels of 458 and 451 m), the indicated subspecies is already dominant among representatives of the genus *Charlesdowniea*, although it coexists with *Ch. coleothrypta* even here. The appearance level of abundant *Wetzeliella articulata*, the index species of the next zone, defines the upper limit of biostratigraphic unit under consideration. Above the lower boundary, *Ch. coleothrypta* appears in association with *Deflandrea phosphoritica, Ch. tenuivirgula, Cribroperidinium tenuitabulatum, Dapsilidinium pastielsii, Melitasphaeridium asterium*, and *Achilleodinium biformoides*. In the middle part, the associated newcomers are *Thalassiphora pelagica, Lejeunecysta hyalina*, and *Areoligera medusettiformis*. Dominant in the zone are representatives of genera *Deflandrea, Cordosphaeridium, Spiniferites, Glaphyrocysta*, and *Wetzeliella*. Species of genera *Systematophora, Spinidinium*, and *Horologinella* are also abundant in some samples. The abundance rate of acritarchs and prasinophytes is insignificant.

The stratigraphic range of *Ch. coleothrypta* dinocyst assemblage is easily recognizable owing to the mass abundance of its index species and subspecies *Ch. coleothrypta rotundata*, which occur in association with diverse representatives of genera *Deflandrea, Spiniferites, Cordosphaeridium*, and *Operculodinium*. *Wetzeliella articulata* is not infrequent as well, but *Alterbidinium* sp. 1, *Areoligera senonensis*, and *Kallosphaeridium* forms characteristic of older assemblages do not cross the lower boundary of *Charlesdowniea coleothrypta* Zone. The occurring single specimens of *Dracodinium varielongitudum*, the index species of underlying zone, disappears near the upper boundary of the zone.

In the core section 011BP, Kul'kova identified at the depth of 492,5 m the following taxa: *Areoligera* cf. *senonensis, Cerodinium sibiricum, Gonyaulacysta* sp., *Deflandrea dissoluta, D. oebisfeldensis, Deflandrea*

sp., *Dracodinium varielongitudum, Deflandrea phosphoritica, Wetzeliella* cf. *articulata, Wetzeliella* sp., *Cordosphaeridium funiculatum, Homotryblium abbreviatum, Homotryblium* sp., *Spiniferites* spp., *Hystrichosphaeridium* sp., *Enneadocysta arcuata, Thalassiphora pelagica, Wilsonidium* sp., *Apectodinium* sp., *Phthanoperidinium stockmansii, Rhombodinium coronatum, Hystrichosphaeridium tubiferum, Samlandia chlamidophora, Hystrichokolpoma* sp., *Achomosphaera* sp., *Spiniferites ramosus, Soaniella granulata*, and *Charlesdowniea tenuivirgula.*

In opinion of Zaporozhets, the upper part of the upper Lyulinvor Subformation in the core section 011BP is already beyond the *Wetzeliella articulata* (acme)–*Systematophora placacantha* Zone of the lower Lutetian. Her opinion is substantiated by absence of index species *Ch. coleothrypta* in samples from the depth levels of 459 and 456 m, while *Wetzeliella articulata* is a common component in the former and dominant taxon in the latter.

The lower boundary of *Wetzeliella articulata* (acme)–*Systematophora placacantha* Zone (Borehole 9, depth interval of 449–444 m; Borehole 011BP, depth interval of 459–456 m) marks the mass appearance level of the first species. This taxon represents up to 15% and over 25 and 10% of dinocyst assemblages from the depth levels of 449.6, 448.5, and 447 m, respectively. Since the greater amount of *Wetziella* forms have not been identified at the species level because of a poor preservation state, the total abundance rate of specimens representing this genus is not less than 50% and more. The second species *Systematophora placacantha* is less abundant and represents 3 to 5% of the assemblages. Its last occurrence is recorded in the green clay member, above which the species does not occur. Nevertheless, the mass development of prasinophytes and acritarchs, but not the disappearance of *Systematophora placacantha* and appearance of single *Areosphaeridium diktyoplokum* species, is the most significant biotic event that has been probably related to a quick shoaling and partial freshening of the basin. As we believe, the upper boundary of the zone should be exactly at this level to be significant in both the stratigraphic (in core section 9) and ecologic aspects. In the core section 011BP, an upper part of the zone corresponds to a break in sedimentation.

The *Wetzeliella articulata–Systematophora placacantha* assemblage is very diverse (70–80 taxa), with well represented chorate cysts (genera *Cordosphaeridium, Operculodinium*, and others). It includes many new taxa (*Glaphyrocysta exuberans, Diphyes ficusoides, Dapsilidinium pseudocolligerum, Homotryblium oceanicum, H. tenuispinosum, Deflandrea arcuata*, and others) in addition to species widespread in older assemblages, such as *Charlesdowniea* forms, the index species of preceding zone included, *Eatonicysta ursulae, Wetzeliella meckelfeldensis, Heteraulacacysta pustulata, Glaphyrocysta ordinata, Apteodinium* sp., *Areoligera* sp., and *Apectodinium* sp. Also characteristic of the zone are very diverse species of genera *Homotryblium, Deflandrea, Cordosphaeridium, Glaphyrocysta, Operculodinium, Spiniferites, Achomosphaera, Apteodinium, Wetzeliella, Thalassiphora*, and *Spinidinium.*

In the core section 9, the uppermost part of the green clay member (depth interval of 444–435 m), which bears fish remains, is attributed to the *Paucilobimorpha–Micrhystridium* Beds of the terminal Lutetian–?basal Bartonian. In the core section 011BP, this interval corresponds to a hiatus. As is noted above, the assemblage with prasinophytes and acritarchs indicative of progressive basin shoaling is primarily of ecologic rather than of age significance, and we withhold comments here about its correlation with the dinocyst zonation.[1] When abundance and diversity of prasinophytes and acritarchs grew, the dinocyst species simultaneously decreased in number to 30–40 euryhaline forms (reduced to a half at least), while siliceous plankton became extinct. *Paucilobimorpha triradiata* and to a lesser extent *P. granuligera* are dominants of the prasinophytes–acritarch assemblage. Both taxa appear at this level for the first time. They coexist with single specimens of *Cymatiosphaera* forms, *Paucilobimorpha tripus*, and *Schizocysta* sp. *Deflandrea* species are as abundant and diverse as in the previous assemblage, but *Wetzeliella* forms are replaced by representatives of the genus *Kisselovia.* Many forms, especially the Paleocene taxa *Cerodinium markovae, C. diebelii, Kallosphaeridium*, and *Deflandrea oebisfeldensis*, are redeposited. Remarkable among species occurring in the beds under consideration are *Wetzeliella echinulata, Lejeunecysta globosa, Glaphyrocysta semitecta, Lentinia wetzelii*, and *Samlandia chlamidophora.* Many other taxa, e.g., *Tectatodinium, Cordosphaeridium, Deflandrea, Apteodinium, Cribroperidinium, Microdinium, Thalassiphora, Phelodinium, Spinidinium, Achilleodinium, Impagidinium, Hystrichokolpoma*, and *Hystrichosphaeropsis*, are of wide stratigraphic ranges.

Planktonic foraminifers. Species *Pseudohastigerina wilcoxoensis* and *Subbotina eoc enica* are identified in two samples from depth levels of 477 and 451 m of the core section 9, i.e., in Member "a" of light gray opoka-like clays. In the core section 011BP, a sample from the borehole bottom (depth 498 m) yielded *Subbotina eocaenica, Pseudohastigerina wilcoxensis, Subbotina pseudoeocena*, and *Acarinina* sp. (determinations by Krasheninnikov and Beniamovskii). In opinion of the latter, representatives of the genus *Acarinina* may be classed with *A. convexa.*

Planktonic foraminifers have been also macerated from two samples (depth levels of 449 and 445 m) characterizing Member "b." At the former level, Beniam-

[1] The terminal beds of the upper Lyulinvor Subformation in the depth range of 440–435 m seem to be corresponding, like the basal Tavda beds, to the *Areosphaeridium diktyoplokum–Rhombodinium draco* Zone, because both index species have been found in a sample from the depth of 440 m.

ovskii identified *Subbotina subtriloculinoides, S. ratusa, S. boweri, Acarinina rotundimarginata* (Krasheninnikov approved identification of the last taxon), *Acarinina* spp., and *Pseudohastigerina micra*, and this assemblage can be attributed to the *Acarinina rotundimarginata* Zone. At the next level, he identified *Hantkenina australis, Acarinina rotundimarginata, A. rugosoaculeata, Subbotina inaequispira, S. pseudoeocaena*, and a fragment of test resembling *S. turcmenica*. In the same sample, Krasheninnikov established presence of *Hantkenina alabamensis* and *Acarinina rotundimarginata*. The entire assemblage can be regarded as characteristic of the *Hantkenina alabamensis* Zone (the upper *H. australis* Subzone) of the upper Lutetian–lower Bartonian.

Benthic foraminifers. Agglutinated benthic foraminifers, among which *Bolivinopsis spectabilis* is leading taxon, have been encountered at the depth of 491 m in Borehole 9. Within the depth interval of 479–443 m, their assemblage is represented by *Ammomarginulina deflexa, A. spectata, Bathysiphon eocenicus, Bolivinopsis spectabilis, Gaudryinopsis subbotinae, Hyperammina friabilis, Hypperamminoides crumena, Rhabdammina cylindrica, Rhizammina indivasa, Labrospira granulosa, L. honesta, Proteonina* spp., *Psammosphaera laevigata, Reophax ampulacea, R. difflugiformis, R. subfusiformis, R. scorpirus, Saccammina sphaerica, Textularia carinatiformis, Trochammina gracilis, T. infirma*, and *Verneulinoides paleogenicus*. The index species of *Textularia carinatiformis* Zone defines biostratigraphic range of this assemblage. Within that zone (depth range from 454 to 446.5 m), there are the *Gaudryinopsis subbotinae* Beds. An assemblage of secretory foraminifers characteristic of the *Uvigerina costellata* Zone was found at the depth of 449 m. This assemblage includes *Bolivina microplancetiformis, U. castellata, Cibicides westi, Nodosaria ewaldi, N. longiscata*, and *Robulis inornata*. Secretory foraminifers from a higher level of the green clay member (depth 445 m) are represented by *Clavulinoides szaboi, Vaginulinopsis fragaria, Heterolepa eocaenica*, and *Planulina costate*, which belong to the upper part of the same *Uvigerina costellata* Zone distinguished in the Lutetian Kuberle and Kresty sequences of European Russia. In the core section 011BP, the entire upper Lyulinvor Subformation corresponds to the *Textularia carinatiformis* Zone characterizing the Lyulinvor sediments according to conclusions of É.N. Kisel'man.

Radiolarians. Kozlova who studied radiolarians from the upper Lyulinvor Subformation of both core sections discriminated three biostratigraphic subdivisions.

(1) The *Heliodiscus inca* Zone is established in the lower part of the upper Lyulinvor Subformation: Borehole 9, depth interval of 478,8–491m, and Borehole 011BP, depth interval of 493–498 m. A typical early Eocene (Ypresian) assemblage of the West Siberian type is confined to the indicated intervals. In addition to the index species, the assemblage includes *Acropyramis magnifica* (Clark et Campbell), *Spongocyclia composita* (Kozlova), *Stylosphaerella irinae*, (Lipman), *Heliodiscus hexasteriscus* Clark et Campbell, and new species *Apoxprunum curganica* Kozlova, which prevail over somewhat less abundant *Conoactinomma stilliformis* (Lipman) and *Lophosphaena sibirica* Gorbovetz. This assemblage of the middle Ypresian age is correlative with the nannoplankton Zone NP12 of the Paleogene standard and with the lower part of the *Morozovella aragonensis* Zone discriminated in succession of planktonic foraminifers of the Crimea–Caucasus region [8].

(2) The *Thecosphaerella turcmenica–Phacodiscinus testatus testatus* Beds span the middle and upper intervals of the upper Lyulinvor Subformation: Borehole 9, depth interval of 458–468.5 m, and Borehole 011BP, depth interval of 481–496 m. The beds reveal a considerable change in taxonomic composition of radiolarians whose assemblage includes a group of transit early Eocene and even late Paleocene taxa, such as *Amphicarydiscus biconstrictus* (Lipman), *Axoprunum inclarum* (Krasheninnikov), *Clathrocyclas multiplicatus* (Lipman), *Eucyrtidium striata exquisite* (Kozlova), and some others. These forms and associated fragments of the Late Cretaceous taxa point to erosion of older sediments, fossil fauna of which is redeposited. At the same time, the beds yield radiolarians, which are characteristic of younger *B. clinata–B. longa* and higher zones known in other regions (areas northward of the Caspian Sea, Ul'yanovsk region of the Volga River basin, and oceanic bottom sediments). These are *Albatrossidinium litos* (Clark et Campbell), *Amphymenium splendiarmatum* Haeckel, and *Periphaena splendida* (Lipman). The other species, *Axoprunum chabakovi* (Lipman), *Phacodiscinus testatus testatus* (Kozlova) and *Thecosphaerella turcmenica* (Lipman) included, which may occur sporadically in underlying horizons, show their mass abundance exactly in the *Buryella clinata–Buryella longa* and *Lychnocanium separatum* zones. Accordingly, the above radiolarian assemblage suggests that the stratigraphic interval under consideration may correspond to the *B. clinata–B. longa* Zone. Nevertheless, the index species and some other taxa characteristic of the zone have not been found in the assemblage, and the interval is attributed therefore to the *Thecosphaerella turcmenica–Phacodiscinus testatus testatus* Beds. The *B. clinata–B. longa* Zone of the terminal Ypresian is correlative with the nannoplankton Zone NP13 and with the upper part of *Morozovella aragonensis* Zone [8].

(3) The *Conocaryomma aralensis–Phacodiscinus testatus subsphaericus* Beds are recognized in the upper and terminal parts of the upper Lyulinvor Subformation: Borehole 9, depth interval of 445–458 m, and Borehole 011BP, depth interval of 459–473,5 m. Their lower boundary is defined at the appearance level of *Acanthosphaera formosa* Krasheninnikov. *Conocaryomma aralensis* Lipman and *Phacodiscinus testatus sub-*

sphaericus (Lipman) appear above this level, and mass accumulations of the latter are very characteristic of the *Lychnocanium separatum* Zone. Nearly at the top of the beds, there have been found *Phacodiscinus fragilis* (Totschilina) and *Stylosphaerella megaxyphos* (Clark et Campbell), the first appearance of which is usually confined to the *L. separatum* Zone. The *Lychnocanium separatum* Zone spans the lower part of the Lutetian Stage. It is correlative with the nannoplankton zones NP14 and NP15 and with *Acarinina bullbrooki* and *A. rotundimarginata* zones of planktonic foraminifers.

According to data of Vitukhin, radiolarians are abundant and diverse at the depth level of 462 m. Nevertheless, some nasselarian species represented by single specimens at the depth of 468.5 m have not been found at the level in question that marks the first appearance of *Siphocampe minuta, Lithostrobus picus, Dictyopora urceolus, Theocyrtis* sp., *Heliodiscus inca*, and *Phacodiscinus* aff. *testatus*. Species *Spongodiscus americanus* and *Spongotrochus radiatus* dominate here among spumellarians, and *Siphocampe minuta* is a dominant taxon of nasselarians.

Radiolarians from the depth range of 458.5–454 m belong most likely to the *Buriella clinata–B. longa* Zone. As the index species is absent in the assemblage, the zone can be discriminated conventionally based on present characteristic taxa *Acanthosphaera formosa* and *Phacodiscinus testatus subsphaericus*. In general, radiolarians are noticeably less diverse in the above interval that yields, in addition to the aforementioned taxa, only *Heliodiscus hexasteriscus, Spongodiscus americanus, Spongotrochus alveatus, S. radiatus, Amphisphaera coronata, Haliomma triglobulata, Eusiringium striata striata, Clathocyclas elegans, Lychnocanium* sp., and *Siphonocampe minuta*.

The assemblage from the depth range of 451–445 m (Member "b") is of a different composition. Species *Cenocaryomma aralensis* and *Calocyclas asperum* found here are characteristic of the *Lychnocanium separatum–Heterestrum schabalkinii* Zone, the index species of which have not been found however. Besides the fragmented Sphaeroidea tests and single nasselarians, the other associated taxa are *Patagospyris* sp., *Petalospyris* sp., and *Lithomelissa* sp.

Diatomaceous algae. In Borehole 9, diatoms are present in the middle and upper intervals of the lower member of the upper Lyulinvor Subformation (depth range of 483–458.5 m), which is composed of light gray opoka-like clays (samples from the depth levels of 483, 478.9, 477, 476, 474, 468.5, 466, 464, 462, 460, and 458.5 m). Diatoms occur as well in the upper member of green clays bearing fish remains (samples from the depth levels of 446, 445.7, and 445.1 m). An impoverished assemblage of neritic diatom genera *Paralia* and *Arachnoidiscus* is characteristic of the depth levels 483 and 478.9 m. Species *Pyxilla gracilis, Grunowiella gemmata*, and *Coscinodiscus payeri* are found at the depth of 477 m. Beginning from the depth of 474 m, diatoms are more diverse. Taxa appearing here are *Coscinodiscus decrescens, C. decrescenoides, Hemiaulus elegans, H. polymorphus* var. *Pseudopodosira westi*, and silicoflagellates *Dyctiocha transitoria* and *Naviculopsis minor*. Newcomers *Coscinodiscus polyactis, C. bulliensis, C. heteroporus, Brightwellia hyperborea*, and *Pseudotriceratium* forms are recorded in the depth range of 462.5(458.5 m, where silicoflagellates *Dyctiocha rotundata* and *D. transistoria (=Naviculopsis minor)* are more abundant than below and representatives of the genus *Ebriopsis* appear as well. According to composition of diatoms, the depth interval of 478.5–458.5 m corresponds to the *Pyxilla gracilis* Zone, and its upper part (460–458.5 m) can be regarded as the *Coscinodiscus polyactis* Beds transitional to the *Pyxilla oligocaenica* Zone. Based on her own zonal scale, Glezer distinguished in the depth range of 466–458.5 m (samples from the depth levels of 466, 462, 460, and 458.5 m) the *Brightwellia hyperborea* assemblage and silicoflagellates of the *Dictyocha secta* Zone, considering them as corresponding to the upper part of the *Pyxilla gracilis* s. l. Zone.

Diatoms and silicoflagellates have not been found in the depth range of 457–446 m, but higher in the section Radionova who studied three samples from depth levels of 446, 445.7, and 445.1 m identified abundant *Pyxilla gracilis* Jouse var. *tenuis* and *Pyxilla oligocaenica* var. *oligocaenica*, which associate with single specimens of *Coscinodiscus decrescens* and *C. payeri*. In the same samples, she found neritic (*Paralia, Pseudopodosira westii, Pyxidicula crenata*) and benthic (*Arachnoidiscus ehrenbergii, Istmia enermis*, and others) diatoms, resting spores, and sponge spicules. The encountered silicoflagellate species are *Mesocena oamaruensis* and *M. eodentalis*. The diversity of diatoms decreases upward, and above the depth level of 445 m, they do not occur. As Radionova concluded, the studied interval of green clays, which accumulated during regression under conditions of the progressing shoaling, corresponds to the upper part of *Pyxilla oligocaenica* Zone. Strelinikova who also studied microfossils from the last depth level arrived at the same conclusion. In opinion of Glezer, samples from the level under consideration characterize the terminal part of *Pyxilla oligocaenica* var. *tenuis* Zone. Until the recent time, Glezer attributed this part to the *Stephanopyxis crenata* Subzone, but now she thinks that it may be regarded as an individual zone. The diatom assemblage from the depth of 443.6 m is unusual, containing species of the genus *Cosmiodiscus*, which are characteristic of stratigraphic levels above the *Pyxilla oligocaenica* var. *tenuis* Zone.

Strel'nikova also investigated diatoms in nine samples from the depth range of 492–458 m of Borehole 011BP. At the depth levels of 492, 490.8, 487.5, and 479 m, she encountered the half-decomposed diatom frustules of unclear taxonomic affinity. Macerates from the depth levels of 475.3, 469.3, 463.3, and 458 m turned out to be most rich in diatoms. An assemblage from the depth of 475.3 m undoubtedly belongs to the

Pyxilla gracilis Zone, and that from the depth interval of 469.3–458 m to the *Pyxilla oligocaenica–P. oligocaenica* var. *tenuis* Zone. Besides the index species, taxa characteristic of the *Pyxilla gracilis* Zone are *Hyalodiscus radiatus, Hyalodiscus* sp., *Pseudoposira westii, Paralia sulcata*, less frequent *Coscinodiscus* sp. aff. *argus, Stephanopyxis megapora*, and single specimens of *Stephanopyxis arctica, Odontotropis carinata, Gyrodiscus vortex, Epithelion rossicum*, and *Hemiaulus* sp. The silicoflagellate group includes dominant *Naviculopsis constricta* and *Dictyocha fibula* occurring in association with less abundant *Dictyocha rotundata* and *D. transitoria*. Ebriids are represented by *Pseudoammodochium dictyoides* and *Ebriopsis exigua*. Three other samples characterize the representative assemblage of the younger zone. At the depth of 469.3 m, the assemblage includes abundant *Pyxilla gracilis, P. oligocaenica, Hyalodiscus* sp., *Pseudopodosira corolla, P. westii*, and *Paralia sulcata*, which coexist with less abundant *Paralia crenulata, Stephanopyxis turris* var. *intermedia, St. turris* var. *cylindrus, St. megapora, Coscinodiscus decrescens, Coscinodiscus* sp. (with sections), and *C. heteroporus*. Ebriids *Ammodochium speciosum, Pseudoammodochium dictyoides*, and *Ebriopsis exigua* are also abundant here. The encountered silicoflagellate species are *Mesocena elliptica, Naviculopsis constricta*, and *Dictyocha rotundata*. Most abundant at the depth of 463.3 m are *Pyxilla oligocaenica, P. oligocaenica* var. *tenius* (slightly less frequent than the type form), *P. gracilis*, and *Hyalodiscus* sp. The associated taxa are represented by less abundant *Pseudoposira westii, P. wittii, P. aspera, Paralia clavigera*, and by single *Anaulus weiprechtii* and *Aulacodiscus* sp. Prevailing among ebriids are *Ebriopsis exigua, E. mesnili, E. aplanata*, and *Pseudoammodochium dictyoides*, whereas silicoflagellates are represented by *Dictyocha elata, D. rotundata, D. obliqua, Mesocena elliptica*, and *Naviculopsis constricta*. Sponge spicules are the main organic remains at the depth of 458 m. Diatom frustules are not rock-forming here. They mostly represent *Pyxilla oligocaenica* var. *tenuis, Paralia sulcata, Hyalodiscus* sp., and rare or single specimens of *Pyxilla gracilis, Coscinodiscus decrescens, Paralia clavigera*, and *Pseudoposira westii*. Among ebriids, only *Pseudoammodiscus dictyoides* is identified.

Spores and pollen. Palynomorphs are insignificant components of organic remains from the upper Lyulinvor Subformation, especially from its lower interval. Consequently, they may be of a corrective significance only by the age determination of particular levels. Aleksandrova who studied palynomorphs from the core section 9 found only scarce pollen grains of conifers (*Pinus, Protocedrus*) and tricolporate form taxa in the depth range of 492.5–483 m. At the depth of 480 m, occasional pollen grains of ancient Pinaceae occur in association with sporadic pollen of Myricaceae, *Plicapollis* ex gr. *pseudoexcelsus*, and *Triporopollenites robustus*. Kul'kova identified *Platycaryapollis pseudocyclus, Plicapollis pseudoexcelsus, Triporopollenites robustus, Castanea crenataeformis, Interpollis supplingensis, Myrica* sp., *Podocarpus* sp., *Taxodium* sp., *Pinus* sp., and spores Polypodiaceae and *Cyathea* sp. in the same section. She correlated this assemblage with the *Triporopollenites robustus–Plicapollis pseudoexcelsus* assemblage [9]. Zaporozhets who studied samples from the depth range of 478–460 m also established here a low content of pollen representing Pinaceae, *Castanopsis pseudocingulum, Pompeckjoidaepollenites* sp., *Tricolporopollenites cingulum, åyricaceae, Triporopollenites* sp., and *Trudopollis* sp. With a certain reservation, she correlated them with the assemblage of *Araliaceaepollenites euphorii, Sapotaceoidaepollenites manifestus, Pompeckjoidaepollenites subhercinicus*, and *Castanopsis pseudocingulum*. Presumably attributable to the same assemblage is the diverse pollen of the form taxa *Anacolosidites* sp., *Rhoipites* sp., *Tricolporopollenites* sp. (dominant), *Triporopollenites* sp., *Triatriopollenites* sp., *Trudopollis pompekjii*, and *Plicapollis* sp., which coexist with accessory Pinaceae, Taxodiaceae, and Fagaceae, such as *Castanopsis* sp. (dominant), *Castanea* sp., *Quercus graciliformis*, and *Q. gracilis*. Scarce pollen grains represent in this assemblage the thermophilic plants, such as Palmae, *Platycarya* sp., *Rhus* sp., Hamamelidaceae, *Liquidambar* sp. and others. In the light-colored opoka-like clays of both core sections, Kul'kova recognized a more complete pollen spectrum of the indicated zone. The spectrum includes *Castanea crenataeformis, Castanopsis pseudocingulum, Loranthus elegans, Engelhardtia quietus, Subtriporopollenites* sp., *Thiatriopollenites aroboratus, Tricolpopollenites* sp., *Tricolpopollenites liblarensis, Reevesia lubomirovae, Ulmodeipites tricostatus, Subtriporopollenites megagranifer, Hamamelis* sp., *Carya* sp., *Nyssapollenites* sp., *Myrtaceidites* sp., Palmae sp., and *Myrica* sp. An insignificant admixture of coniferous pollen represents *Podocarpus* sp., *Cedrus* sp., *Pinus* spp., and Taxodiaceae. The present spores are those of *Gleichenia* sp., *Osmunda* sp., Polypodiaceae, and *Lycopodium* sp. Some samples yielded pollen of the *Longaxones* type.

In samples from the depth interval of 454–446.3 m of Borehole 9, which have been studied by Zaporozhets, palynomorphs are of the same composition, and their content gradually increases from 10 to 30% upward in the section. Pollen of conifers is insignificantly abundant as before and represents a third of the spectrum at the depth of 447 m only. The form taxa identified in this interval are *Rhoipites* sp., *Tricolporopollenites* sp., *Triporopollenites* sp., *Plicapollis* sp. (possibly redeposited), *Trudopollis* sp., *Tricolporopollenites cingulum*, and *Triatriopollenites myricoides*. Dominant among natural taxa are Fagaceae (*Castanea, Castanopsis*, and evergreen *Quercus* species) in association with rather abundant Hamamelidaceae. This set of taxa may be correlated with the *Castanea crenataeformis–Castanopsis pseudocingulum* assemblage of West Siberia.

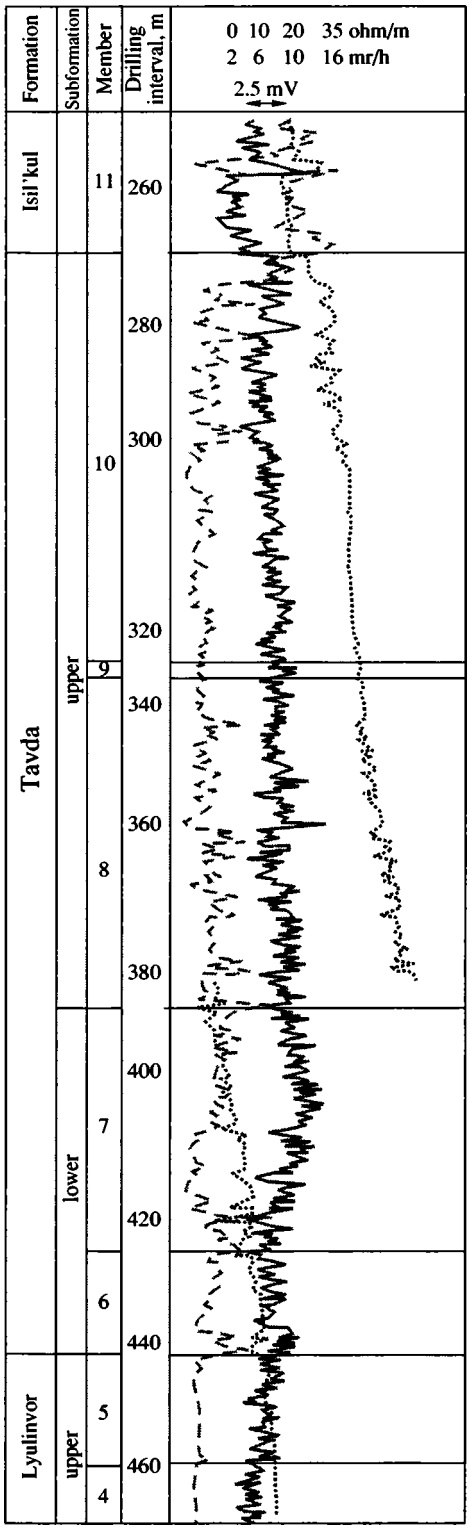


Fig. 7. Electric logs, Borehole 9, depth interval of 460–250 m (symbols as in Fig. 6).

The abundance and composition of palynomorphs sharply change in the section interval between 444 and 435 m, where the basin shoaling is recorded. Spores and pollen represent here 40 to 60% of the spectrum. They considerably prevail over dinocysts, percentage of which is equal to 10%. Pollen of flowering plants is somewhat more abundant than acritarchs and prasinophytes, which are more concentrated at this level as well. Conifers and Fagaceae constitute 25 and up to 50% of the spectrum, respectively. *Castanea* and *Castanopsis* (*C. pseudocingulum*) are still present among Fagaceae, but evergreen *Quercus gracilis* and *Q. graciliformis* are dominant in this family. Juglandaceae and Hamamelidaceae represent up to 8–9% of palynological spectra, while Betulaceae occur as occasional specimens. The identified Juglandaceae taxa are *Pterocarya* (up to 4%), *Juglans* (up to 3%), *Engelhardtia, Platycarya*, and *Carya* (scarce pollen grains). In the family Hamamelidaceae, percentages of *Corylopsis, Hamamelis, Liquidambar*, and *Fothergilla* (*F. gracile* included) are approximately equal (about 2%). The other taxa occurring in perceptible amounts are *Nyssa* (4%), *Rhus* (2%), *Acer* and *Comptonia* (1%). Pollen of *Myrica, Ulmus, Planera, Zelkova*, Moraceae, Ericaceae, *Tilia, Ilex, Eleagnus*, Oleaceae, *Cornus*, and *Nelumbo* is represented by single grains. Form species representing in sum not more than 5% are *Tricolporopollenites* sp. (4%), *Trudopollis pompeckjii, Plicapollis* sp., *Rhoipites* sp., and *Triporopollenites* sp. The described assemblage can be correlated with the regional *Quercus gracilis–Castanopsis pseudocingulum* assemblage.

Magnetostratigraphy. Seven magnetic polarity zones established in the upper Lyulinvor Subformation of the core section 9 are as follows: the normal polarity zones N_3ll, N_4ll, N_5ll, and N_6ll, and the reversed polarity zones R_3ll, R_4ll, and R_5ll with the normal polarity subzones inside two of them (Fig. 5). The zones range in thickness from 3 to 14 m. Two zones R_1ll and N_1ll, which are 20 and 22 m thick respectively, are established in this subformation recovered by Borehole 011BP.

The boundary between *Dracodinium varielongitudum* and *Charlesdowniea coleothrypta* is at the depth of 493.3 m inside the basal reversed-polarity zone R_1ll of the core section 011BP and at the depth of 466 m within the reversed polarity zone R_4ll in Borehole 9. As for the lower part of *D. varielongitudum* Zone penetrated by Borehole 9 only, it spans the interval of polarity zones N_3ll, R_3ll, N_4ll, and R_4ll (lower part). In the upper part of subformation recovered by Borehole 011BP, the *Charlesdowniea coleothrypta* and *Ch. coleothrypta rotundata* zones correspond to the interval spanning the greater part of the reversed polarity zone R_1ll and the normal polarity zone N_2ll with two reversed polarity intervals inside it. In the core section 9, the same stratigraphic interval corresponds to the reversed polarity zone R_4ll in its greater part, normal polarity zone N_5ll, and reversed polarity zone R_5ll. Frequent erosion

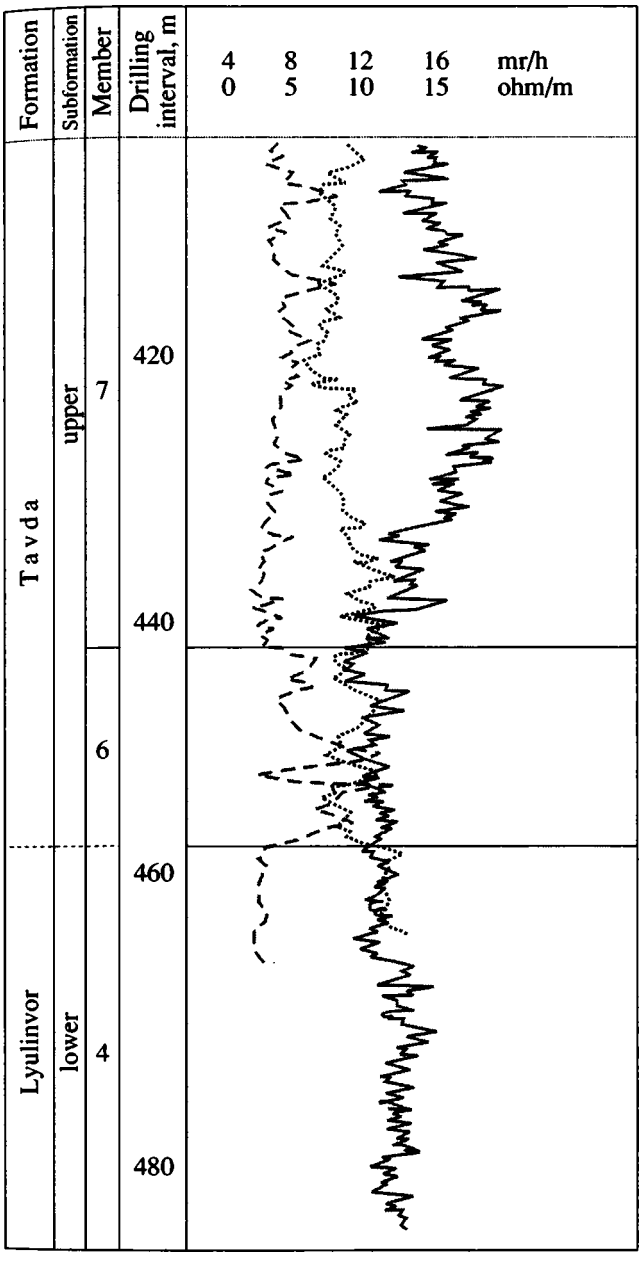


Fig. 8. Electric logs, Borehole 011BP, depth interval of 460–400 m (symbols as in Fig. 6).

events recorded in the upper Lyulinvor subformation and decreased thickness of magnetostratigraphic units are the main obstacles for the correct interpretation of paleomagnetic data. It is very likely that the upper Lyulinvor Subformation was partially eroded before the Tavda time as it follows from palynological and paleomagnetic data (decreased ranges and eliminated fragments of magnetic polarity zones).

Electric logging. Electric logs obtained for the upper Lyulinvor Subformation are differentiated even lesser than within the lower subformation. The AR values correspond in average to 3–4 ohm, and SP parame-

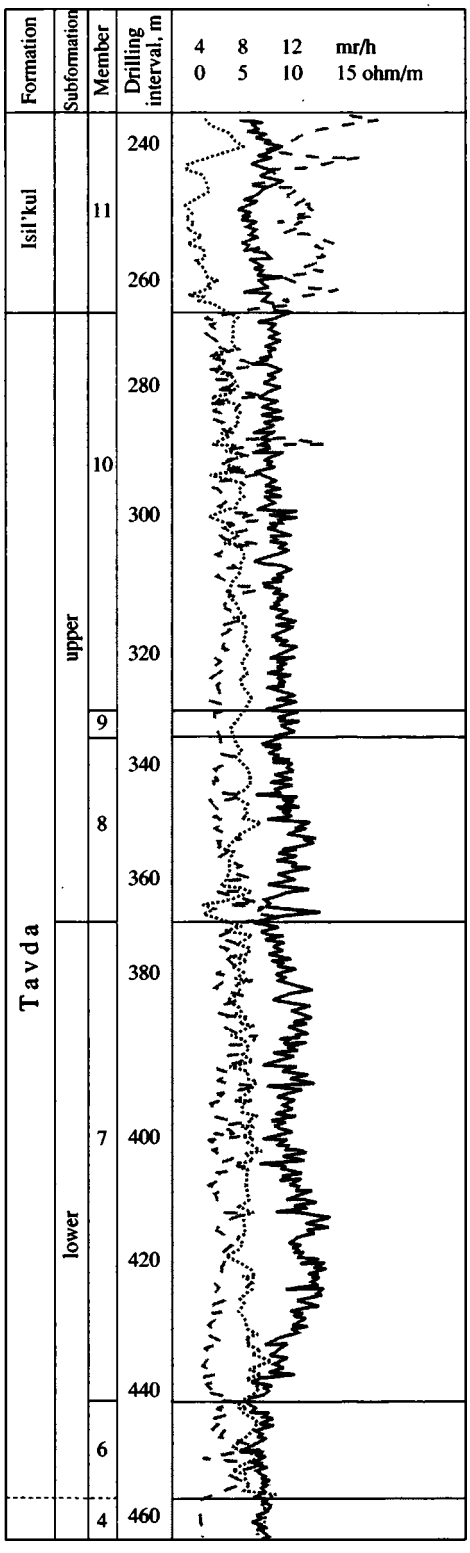


Fig. 9. Electric logs, Borehole 011BP, depth interval of 460–240 m (symbols as in Fig. 6).

ters range from 3 to 5 mV. The opposite displacements of both parameters are clearly recorded in the log for Borehole 9 at the level of basal bed of glauconite sandstone. The other peaks of AR and SP values are also confined to sandstone bed at the depth level of 470 m. The transition from light gray opoka-like to green nonsiliceous clays is depicted by a minor decrease of AR values at the level of SP parameters slightly displaced to the right (Figs. 6–9).

Tavda Formation

Except for the basal interval, the rest of the Tavda Formation sequence is of the so-called "Kulunda" type, since rocks of this kind are widespread in the synonymous lithologic–facies zone on the south of West Siberia. The thickness values (152 m in core section 9 and 178 m in core section 011BP) are close to the maximal ones known in the Omsk depression and in central areas of the West Siberian plate. The lower and upper intervals of the sequence are composed of clays. In the middle interval 50 m thick in core section 9 and 55 m thick in core section 011BP, there is recorded a transition in the eastward direction from coastal-marine to continental facies enriched in sandy material and hosting the interlayers of brown coal and segregations of hydrogoethite–limonite ores. The lower interval corresponds to the lower Tavda Subformation that is homogeneous in composition, being 47 m thick in core section 9 and 53 m thick in core section 011BP. In the former section, this subdivision is composed of greenish foliated fat clay with rare siderite or marly-siderite nodules and interlayers ranging in thickness from 0.1 to 0.5 m. The basal 10-m-thick member of fine-grained glauconite sandstone and siltstone has a sharp contact with the underlying Lyulinvor Formation. The subformation upper boundary is placed at the depth level of 388 m, which corresponds to the base of silty clays, the main rocks of the middle interval. A thin bed of siderite sandstone is confined to this boundary. In the core section 011BP, the basal clay–aleuritic member of the lower Tavda Subformation is also 10 m thick. Glauconite grains and abundant clasts or pellets derived from the underlying Lyulinvor clay are concentrated in the clayey–silty–sandy matrix near the base of this member. Transition between the lower and middle members is gradual. An interlayer of micaceous siltstone at the depth of 401 m, above which the clay beds are enriched in silty material, can be conventionally regarded as a marker of the lower boundary of the upper Tavda Subformation. In certain intervals of the section, silty clays enclose interlayers of sand, siderite, and pure clay. Leaf impressions of aquatic fern *Azolla vera* are found in both core sections at the depth of 330–331 (Borehole 011BP) and 337 m (Borehole 9). The upper interval predominantly composed of clay encloses frequent siltstone and sandstone interbeds and siderite lenses and interlayers. The boundary between this and middle interval is conventionally defined at the depth of 320 to 330 m in the core section 9 and at the depth of 300 m in the core section 011BP, because the organic-walled phytoplankton found above these levels suggests the more typical marine environments of sedimentation. The Tavda Formation top is eroded, overlain by kaolinic sands of the Isil'kul Formation, which form sole casts and enclose flattened pebbles of the underlying Tavda clay.

Organic-walled phytoplankton. Kul'kova studied 25 to 30 samples of the Tavda Formation rocks from both core sections, and Zaporozhets investigated 18 samples from Borehole 9 and seven samples from Borehole 011BP. The organic-walled phytoplankton of the Tavda Formation developed under a strict control of the sea basin hydrology. Prasinophytes and acritarchs dominate in two intervals: near the base and in the middle of the formation sequence. In the basal sandy member that grades into overlying silty clays, representatives of the late Lyulinvor assemblage are dominant. The middle interval rich in silty material characterizes sedimentation under conditions of shoaling and seawater stratification, as we believe, with a lower salinity in the upper part of the photic zone. This is evident from abundance of aquatic fern spores (*Hydropteris, Azolla*) and from presence of *Pediastrum* remains, the algal taxon characteristic of fresh waters. Both transgressive clay-rich members most characteristic of the lower and upper Tavda subformations yield the diverse dinocyst assemblages (30 to 40 and more species). The diversity of dinocysts considerably decreases down to 10–15 species within the intervals dominated by prasinophytes and acritarchs (Table 2).

Three dinocyst assemblages characterizing compositions of successive biostratigraphic zones are distinguished in the studied core sections (they are especially representative in core section 011BP). These are the assemblages of (1) *Areosphaeridium diktyoplokum–Rhombodinium draco* Zone of the lower Bartonian, (2) *Rhombodinium porosum* Zone of the upper Bartonian, and (3) *Charlesdowniea clathrata angulosa* Zone of the Priabonian, the composition of which is locally complicated by superimposed anomalous assemblages rich in prasinophytes, acritarchs, and spores of aquatic ferns. We also propose to discriminate the middle interval of the formation in the rank of *Hydropteris–Pediastrum* Beds (in the dinocyst zonation, this interval corresponds to the *Charlesdowniea clathrata angulosa* Zone, because single specimens of this index subspecies have been found in the indicated beds at the depth of 325 m in the core section 011BP).

In the core section 9, the *Areosphaeridium diktyoplokum–Rhombodinium draco* Zone spans the depth interval of 435–408 m, where the first index species has been found at all the sampling levels. Zaporozhets established the first occurrence level of the second species at the depth of 408 m near the zone upper boundary and detected its single specimens in the overlying zone. According to data obtained by Kul'kova, this taxon is

present in all the samples from the lower Tavda Subformation beginning from its basal bed. In Borehole 9, the upper boundary of the zone is at the depth of 408 m corresponding to the appearance level of *Rhombodinium porosum*, the index species of the next zone.

The *Areosphaeridium diktyoplokum–Rhombodinium draco* Zone recorded considerable changes in composition of dinocysts. Many species characteristic of the first half of the Paleogene disappear at this level marking the appearance of many newcomers. Among the latter, there are *Thalassiphora elongata* and *Wetzeliella irtyschensis*, the endemics of West Siberia, and *Thalassiphora pelagica, Samlandia reticulifera, Homotryblim plectilum, Spiniferites wetzelii, Pterodinium* sp., *Corrudinium incompositum*, and others. Many newcomers, e.g., *Microdinium reticulatum, Adnatosphaeridium robustum, Wetzeliella articulata?*, and *Charlesdowniea tenuivirgula*, pass into the next zone. The transit species are *Samlandia chlamidophora, Palaeocystodinium golzowense, Deflandrea phosphoritica, Glaphyrocysta semitecta*, and *Spiniferites ramosus*. Redeposited forms (species of genera *Cerodinium, Alterbidinium*, and others) are characteristic of basal beds of the formation, which have been recovered by both boreholes. Species *Paucilobimorpha triradiata* are abundant in the basal sandstones.

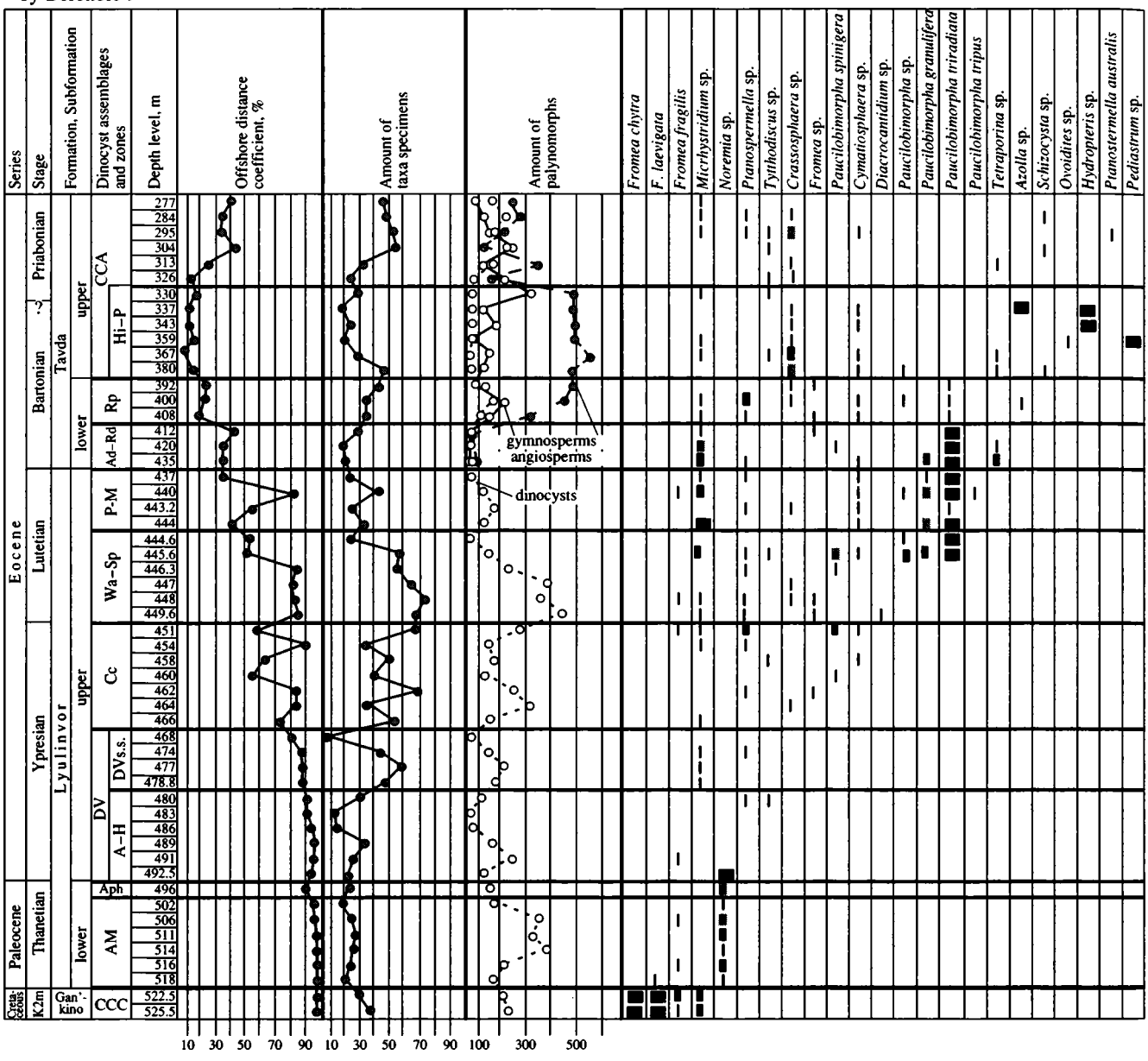
The *Rhombodinium porosum* Zone characterizes the upper part of the lower Tavda Subformation in the depth interval of 408–388 m of Borehole 9. Its lower boundary is at the first occurrence level of the index species, and the upper one is marked by disappearance of taxa characteristic of the zone and by appearance of rather abundant spores of aquatic ferns in association with thin-walled *Hystrichokolpoma* forms, prasinophytes, and acritarchs, which define the lower boundary of the *Hydropteris indutus (Pediastrum)* Beds. *Charlesdowniea clathrata angulosa*, the index subspecies of the next zone appears at the same level. In the core section 011BP, species *Rhombodinium porosum* have not been encountered, and it is difficult to discriminate here the zone in question from the underlying one.

Characteristic of the *Rhombodinium porosum* Zone are also some taxa, which do not occur in the overlying strata. These are *Wetzeliella irtyschensis, Kisselovia ornata, Homotryblium oceanicum, H. abbreviatum, Thalassiphora elongata, Achilleodinium biformoides, Glaphyrocysta undulata, Diphyes colligerum*, and *Charlesdowniea tenuivirgula*. The development of *Phthanoperidinium geminatum, Pht. eocenicum, Cordosphaeridium exilimurum, Chytroesphaeridia* sp., and some other forms commenced in the West Siberian basin in this zone only. About a half of taxa (*Charlesdowniea clathrata, Lentinia serrata, Selenopemphix nephroides, Diphyes colligerum, Palaeocystodinium golzowense, Membranophoridium aspinatum*, and species of genera *Phthanoperidinium, Deflandrea, Thalassiphora*, and *Cordosphaeridium*) are transit. Kul'kova [10] formerly considered *Kisselovia ornata* or *Wetzeliella irtyschesis* as the index species of the zone, but these two taxa are endemics occurring in West Siberian sections only. In distinction, *Rhombodinium porosum* is the index species of concurrent Eocene zones in European Russia and West Europe, thus being more appropriate for the nomenclature.

The *Hydropteris indutus–Pediastrum* Beds span the depth intervals of 380–330 and 361–322,7 m in boreholes 9 and 011BP, respectively. Dinocysts are rare in the beds, which reveal a fantastic abundance of spores of aquatic fern *Hyrdopteris indutus* (appear at the depth of 380 m and are abundant at the depth levels 359, 343, and 337 m in core section 9 and at the depth of 325 m in core section 011BP). Some samples yielded the *Azolla* spores and remains of fresh-water alga *Pediastrum*. Occasional specimens of green algae *Crassosphaera, Tytthodiscus, Pterospermella*, and acritarchs *Micrhystridium* spp. are present almost in all the samples studied. Dinocysts are commonly represented by thin-walled *Hystrichokolpoma* forms associated with species of genera *Phthanoperidinium, Cordosphaeridium*, and *Deflandrea* and with *Membranophoridium aspinatum, Selenopemphix nephroides*, and some other species. As is noted above, *Charlesdowniea clathrata angulosa*, the type subspecies of the youngest dinocyst zone of the Eocene succession, appears in the upper Tavda Subformation penetrated by Borehole 9 above the *Hydropteris indutus–Pediastrum* Beds. In the core section 011BP, this taxon is dispersed throughout the beds, being dominant near their top (sample from the depth level of 325 m). Accordingly, it is reasonable to consider the *Hydropteris indutus–Pediastrum* Beds as a subdivision of the indicated zone.

The *Charlesdowniea clathrata angulosa* Zone is over 100 m thick in the core section 9 (depth interval of 380–273 m) and even thicker in the core section 011BP (?361–265 m). Recognition of its lower boundary is complicated by the reduced diversity of dinocyst species in the *Hydropteris indutus–Pediastrum* Beds. In the core section 9, this boundary is conventionally placed at the disappearance level of *Wetzeliella irtyshensis* and *Kisselovia ornata* characteristic of the *Rhombodinium porosum* Zone, while in the core section 011BP it is defined at the first occurrence level of the index species. The taxonomic characterization of the zone is clear despite the broad sampling intervals. The index species coexists here with characteristic *Phthanoperidinium* species, such as *Phth. stockmansii* and *Phth. geminatum*, and with *Membranophoridium aspinatum, Rottnestia borrusica, Thalassiphora delicata, Enneadocysta arcuata*, and *Deflandrea phosphoritica* f. *attenuata*. Species *Araneosphaera araneosa, Phthanoperedinium resistente, Selenopemphix selenoides, S. coronata*, and others appear beginning from the basal level of *Hydropteris indutus–Pediastrum* Beds. Taxa successively appearing in the middle and upper intervals of the zone are *Heteraulacacysta porosa, Glaphyrocysta vicina, Lejeunecysta cinctoria, Batiacasphaera sphaerica, Phthanoperidinium amoe-*

Table 2. Distribution ranges of acritarchs, prasinophytes, and spores *Azolla vera, Hydropteris indutus* in the section penetrated by Borehole 9



num, Phth. flebile, Phth. alectrolophum, Trithyrodinium sp., *Charlesdowniea clathrata angulosa, Ch. variabilis, Hystrichosphaeropsis camplanata, Pentadinium gonoferum, Operculodinium placitum*, and some others (core section 9). The upper transgressive member of clayey beds (depth range of 295–277 m) recorded the recurrent appearance of species characteristic of lower beds of the lower Tavda Subformation (*Rhombodinium longimanum, Lentinia serrata*, and some others). Dinocysts from this depth range are especially diverse, represented by 50 species. Species *Phthanoperidinium amoenum* appears here earlier than in other regions of West Siberia.

Foraminifers. Beniamovskii analyzed 70 samples from different levels of the Tavda Formation and found foraminifers in 16 samples. The assemblage of planktonic foraminifers from Borehole 9 (depth range from 415 to 400 m) includes *Pseudohastigerina micra, Pseudogloboquadrina primitiva, Catapsydrax martini, C. echinatus, C. dissimilis, Catapsydrax sp., Subbotina pseudobulloides, S. linaperta, Acarinina* cf. *bullbrooki*, and *A. rotundimarginata.* The last taxon defines the Bartonian age of host deposits.

Species *Cyclammina crassa* of benthic foraminifers were found at the depth levels of 415.5, 415, 414.5, and 412 m. The *Cribroelphidium rischtanicum* assemblage with index species and *C. differensaperto* is established

in the depth interval of 339–334 m (sampling levels at 339, 337.6, 337.3, and 334 m). These benthic taxa are characteristic of the Upper Eocene in Central Asia (Kisel'man, 1978).

Osracodes. Ostracode species found at different depth levels are *Clithrocytheridea* sp. juv. (413.5 m), *Loxoconcha* sp. (408 m), and *Cyamocytheridea* aff. *corrugata* (406.5 m). Species *Cytheridea probata, C. testate, Pontocypris sibirica, Loxoconca involuta, L. minsaensis, Loxoconca spp., Cytheromorpha barbantica, Cytheromorpha* sp., *Echinocythereis spongiosa, Clithrocytheredia innae, Bythoceratina* ex gr. *inpressa, Parascypris contracta*, and *Cytherura placida*, which are typical of the upper Eocene in the Aral–Turgai region, characterize the ostracode assemblage from the depth interval of 343–330 m. The transgressive clay member of the upper Tavda Subformation yielded ferruginate casts of *Loxoconcha* sp. and *Cytheridea* sp. (depth level of 291 m).

Spores and pollen. In opinion of Kul'kova, the Tavda Formation spans intervals of two regional palynological zones. These *Castanopsis pseudocingulum–Rhoipites* spp.–*Quercus gracilis* (or *Quercus graciliformis* in terminology of Zaporozhets) and *Quercus gracilis–Q. graciliformis* zones correspond respectively to the lower and upper Tavda subformations. Pollen of angiosperm plants, mostly of Fagaceae (up to 50%) and Hamamelidaceae (8 to 10%), is dominant in palynological spectra of the lower Tavda Subformation. A similar trend in distribution of palynomorphs is recorded in the lower half of the upper Tavda Subformation (*Hydropteris indutus Pediastrum* Beds). Several samples from the depth range of 330–313 m of Borehole 9 showed the anomalously high content (over 75%) of pollen produced by gymnosperm plants. In upper, clearly marine facies of the Tavda Formation, pollen of angiosperm plants is again dominant, though to a lesser extent than in the lower subformation. Palynomorphs of higher plants are usually as frequent as 500 to 600 specimens per a preparation. In the upper clay member of the upper Tavda Subformation, where abundance and diversity of dinocysts are higher (30 to 40% of the entire palynomorph assemblage), pollen of higher plants is less abundant. Pollen of Pinaceae is dominant in the group of gymnosperms, being noticeably more abundant than pollen of Taxodiaceae. Dominant among angiosperms are Fagaceae, while Juglandaceae are less abundant and Betulaceae are accessory. Pollen of thermophilic flowering plants is diverse, but many taxa are represented in pollen spectra by occasional specimens only.

Magnetostratigraphy. Eight magnetic polarity zones, four of normal polarity (N_1tg, N_2tg, N_3tg, and N_4tg) and four of reversed polarity (R_1tg, R_2tg, R_3tg, and R_4tg), all with inner subzones of opposite polarity, are established in the Tavda Formation sequence penetrated by Borehole 9. In the core section 011BP, there are nine zones, five of normal polarity (N_1tg, N_2tg, N_3tg, N_4tg, and N_5tg) and four of reversed polarity (R_1tg, R_2tg, R_3tg, and R_4tg), some of which have inner intervals of opposite polarity. In general, the normal polarity zones prevail over others in both sections of the Tavda Formation (Fig. 5).

Electric logging. Clays of the Tavda Formation, which are enriched in silty material, are characterized in logs by the higher AR parameters (6–8 locally up to 10 ohm). At the levels of sand and sandstone interbeds, this parameter increases up to 20–30 ohm. The SP values can be as high as 15–20 mV, being at the maximum within the depth range of basal member composed of sand and silty clay in both sections. The opposite AR and SP peaks at this depth are especially well seen in the log for Borehole 9. Additional peaks are recorded in both logs at the levels of siderite interlayers and lenses (Figs. 6–9).

CONCLUSION

Two core sections of reference boreholes drilled in the south of West Siberia are studied using a complex of bio- and magnetostratigraphic methods. Paleontology of found organic remains characterizing various groups of microplankton is described in detail. Four discriminated lithostratigraphic subdivisions are the Gan'kino Formation, lower and upper Lyulinvor subformations, and Tavda Formation. The first and second subdivisions have been penetrated by Borehole 9 only.

The Gan'kino Formation is composed of marly clays, whereas the overlying strata of marine Paleogene are all less calcareous.

The lower Lyulinvor Subformation characterizes the main stage of organic silica accumulation. It is composed of opoka, opoka-like clay, sandstone, and siltstone beds. The basal glauconite sandstone of the unit encloses phosphatic nodules and overlies with a hiatus the underlying Gan'kino Formation. The formation corresponds to a separate sedimentation cycle that is incomplete, because its upper layers are eroded.

The upper Lyulinvor Subformation is composed, in distinction from the lower one, of fine-grained clayey–siliceous sediments, which yield the diverse assemblages of microplankton. It is of the two-member structure. The thicker lower member is composed of gray opoka-like clay with a basal bed of glauconite sandstone. The overlying member of green, slightly siliceous clays bears fish remains. A break in sedimentation between two members is inferable based on lithologic features and abrupt compositional changes in microplankton assemblages. Accordingly, it seems reasonable to return into stratigraphic nomenclature the Nyurol'ka Formation that has been originally discriminated by S.B. Shatskii and is abolished now.

The glauconite sandstone bed with phosphatic nodules at the lower subformation base signifies the commencement of new sedimentary cycle after erosion of underlying sediments. Borehole 9 drilled near the

Table 3. Correlated zonal scales for different groups of microfossils, the marine Paleogene sections of West Siberia

Stage	Formation	Subformation	Member	Dinocysts	Radiolarians	Silicoflagellates	Diatoms	Foraminifers: benthic	Foraminifers: planktonic	Spores and pollen	Nannoplankton	Ostracodes	Pelecypods
Priabonian	Tavda	upper	10	*Charlesdowniea clatrata angulosa*						*Quercus gracilis, Q. graciliformis*			
Bartonian			9	*Hydropteris indutus – Pediastrum* sp. Beds				*Cribroelphidium rischtanicum*				*Cytheridea probata, Echinocythereis spongiosa, Paracypris contracta* Beds	*Musculus* cf. *samensis* Beds
			8										
		lower	7	*Rhombodinium porosum*					*Subbotina praebulloides* Beds	*Quercus gracilis, Q. graciliformis, Castanopsis pseudocingulum*		*Cyamocytheridea* aff. *corrugata*	
				Areosphaeridium diktyoplokum, Rhombodinium draco				Cyclammina crassa Beds					
			6										
Lutetian	Lyulinvor	upper	2	*Paucilobimorpha – Mystricrhystridium* Beds						*Quercus gracilis, Rhoipites granulatus, Castanopsis pseudocingulum*			
			5	*Wetzeliella articulata* (acme), *Systemathopora placacantha*			*Cosmiodiscus* Beds						
			1		*Conocaryomma aralensis – Phacodiscinus testatus subsphaericus* Beds		*Stepanopyxis crenata* / *Pyxilla oligocaenica tenis* (terminal)	*Uvigerina costellata*	*Hantkenina alabamensis* (*H. australis* Subzone)	*Castanea crenataeformis, Castanopsis pseudocingulum*			
				Charlesdowniea (s.l.) *coleothrypta rotundata*				*Textularia carinatiformis* / *Gaudryinopsis subbotinae*	*A. rotundimarginata*	*Araliaceoipollenites euphori, Sapotaceoidaepollenites manifestus, Pompeckjoidaepollenites subhercynicus, Castanopis pseudocingulum*			
Ypresian			4	*Charlesdowniea coleothrypta coleothrypta* (s.s.)		*Dyctyocha secta*	*Brightwellia hyperborea* / *Pyxilla gracilis* / *Coscinodiscus polyactis*						
			3	*Dracodinium varielongitudum* s.s.	*Thecosphaerella turcmenica – Phacodiscinus testatus testatus* Beds					*Pompeckjoidaepollenites subhercynicus, Castanopsis pseudocingulum*			
				Dracodinium varielongitudum s.l.						*Triporopollenites robustus, T. excelsus, Plicapolis pseudoexcelsus, Myrica*			
				Alterbidinium sp. 1 – *Hystrichosphaeridium tubiferum* Beds	*Heliodiscus inca*			*Bolivinopsis spectabilis*		Occasional *Pinus* sp., *Protocedrus* sp.			
Thanetian		lower	2	*Apectodinium homomorphum*						*Trudopollis menneri, Tricolporopollenites* sp.			
				Alisocysta margarita				*Glomospira gordialiformis*					
Maastrichtian	Gan'kino		1	*Cerodinium diebelii* Beds				*Spiroplectammina kasanzevi, Bulimina rosenkrantzi*		Assemblage of *Pinus* sp., *P. aralica, Protocedrus* sp. and single *Trudopollis* sp.	CC26 *Nephrolithus frequs*	Assemblage of *Confusina notabilis, Cythere*(?) *sibirica*	

southern flank of the West Siberian sedimentary basin recovered an incomplete sequence of the unit, because the sea transgression reached this site later than the northern and central areas of the West Siberian plate. This is evident from the fact that the upper Paleocene–lower Eocene strata known in the Central and Trans-Uralian structural–lithologic zones are absent in the subformation sequence.

The green clay member yielded two assemblages of planktonic foraminifers, which belong to *Acarinina rotundimarginata* and *Hantkenina alabamensis* zones different in age and have been correlated with concurrent assemblages of dinocysts and siliceous microfossils. The results obtained clarify the member age and suggest that beds bearing the above assemblages of planktonic foraminifers are separated by a hiatus.

Now we can also convincingly argue for a considerable basin shoaling at the accumulation time of terminal beds of the green clay member. The arguments in favor of this event are the increased content of silty material in the clay matrix, on the one hand, and changes in composition of diatom assemblages and the appearance of abundant acritarchs and prasinophytes, especially of *Paucilabimorpha* forms, among the organic-walled phytoplankton, on the other.

The three-member structure of the Tavda Formation, which is characteristic of its sections in the Barabinsk and Kulunda lithologic-facies regions, is confirmed in both core sections. That structure is a consequence of two separate transgressive–regressive cycles, the last of which is reduced and represented by the transgressive part only. These peculiarities of sedimentation influenced the dynamics of palynological spectra. They were responsible for predominance of diverse dinocyst species in the lower and upper clay members, the principal ones in the lower and upper Tavda subformations, and for abundance of aquatic fern spores, which occur in association with prasinophytes and acritarchs in the middle interval that is enriched in silty material and ranked as the *Hydropteris indutus–Pediastrum* Beds.

The lower clay member of the lower Tavda Subformation and its basal bed of glauconite sandstone with clay pellets derived from underlying rocks of the Lyulinvor Formation correspond in range to the *Areosphaeridium diktyoplokum–Rhombodinium draco* and *Rhombodinium porosum* dinocyst zones. The *Acarinina–Catapsydrax* assemblage of planktonic foraminifers is obtained from clay interlayers in the middle interval of the subformation sequence. The upper Tavda Subformation that is silty in the lower part (*Hydropteris indutus–Pediastrum* Beds) and clayey in the upper one is attributed to the *Charlesdowniea clathrata angulosa* Zone. In both core sections, a marker bed rich in remains of macro- and microfauna (mollusks, ostracodes, benthic foraminifers, massulae and megaspores of *Azolla*, mass accumulations of *Hydropteris indutus* spores) is confined to the upper part of the shallow-water silty member.

The regional biostratigraphic scales for various groups of biota are coordinated between each other based on different representatives of microplankton occurring jointly in sediments of the Gan'kino Formation and upper Lyulinvor Subformation (Table 3). The diverse assemblage of nannoplankton found for the first time in the Gan'kino Formation of West Siberia is correlated with the dinocyst zones recognized in the same subdivision. Four discrete levels are established in succession of planktonic foraminifers, and corresponding foraminiferal assemblages are correlated with distribution ranges of siliceous and organic-walled phytoplankton, ostracodes, spores, and pollen. Dinocysts present in all the samples studied are used to define levels of biotic events presumably connected with breaks in sedimentation. The beds with abundant prasinophytes, acritarchs, and aquatic fern spores are indicative of the past shoaling events and seawater stratification in the basin. All the results obtained and age substantiation of discriminated biostratigraphic units will be discussed in the next our publication.

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Geochronology of the Neogene–Quaternary Dacitic Volcanism in the Northwestern Lesser Caucasus (Georgia)

V. A. Lebedev*, I. V. Chernyshev*, M. M. Arakelyants*, Yu. V. Gol'tsman*, O. Z. Dudauri, and G. T. Vashakidze****

**Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Russian Academy of Sciences, Staromonetnyi per. 35, Moscow, 119017 Russia*

***Geological Institute, Academy of Sciences of Georgia, ul. Zoi Ruchadze 1, Tbilisi, 380093 Georgia*

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Abstract—The isotopic geochronology of a representative collection of moderately acidic volcanics from the Dzhavakheti neovolcanic area in the northwest of the Lesser Caucasus (Georgia) is studied using the K–Ar and Rb–Sr methods. Three discrete stages of the regional volcanic activity took place in that area during the late Miocene (about 8–7 Ma), Pliocene (4–2 Ma), and Quaternary (800–about 50 ka). Moderately acidic volcanic rocks were most characteristic of the first and third stages, whereas basalts with subordinate rhyolite–dacitic lavas were typical of the second one. The K–Ar dates evidence for intermittent formation of the Dzhavakheti dacites formerly attributed altogether to one Goderdzi Formation. Pyroclastic units and intraformational dacitic flows exposed in the Kura River valley were formed in the late Miocene (7–8 Ma). The extrusive domes of the Amiranisgora and Chikiani volcanoes emerged at about 3 Ma. Dacitic lavas of the Dzhavakheti Ridge were erupted probably in the late Pliocene (about 2 Ma). Dacitic volcanism of the Samsari Ridge that commenced at about 800 ka had three discrete phases responsible for the formation of the "western range" volcanoes (800–700 ka), the Samsari caldera and other volcanoes of the main ridge (300–200 ka), and parasitic volcanoes and hyalodacitic lava flows (later than 50 ka). The volcanic activity in the Samsari Ridge continued up to the late Neopleistocene or, probably, to the Holocene. Accordingly, this area of the Dzhavakheti region is potentially hazardous with respect to volcanism reactivation. It was found out that plagioclase phenocrysts of the Dzhavakheti dacitic lavas contain the ^{40}Ar excess and differ in initial Sr and Nd isotopic ratios from the rock groundmass. These peculiarities previously noted for Quaternary lavas in other Caucasian volcanic centers appear to be also characteristic of Pliocene rocks. It is confirmed that the K–Ar dating of young volcanics, when applied to their groundmass, yields the most reliable results.

Key words: isotopic dating, geochronology, K–Ar method, Rb–Sr method, Lesser Caucasus, Dzhavakheti area, dacites, Neogene–Quaternary volcanism, stages of volcanic activity.

INTRODUCTION

Geochronology of the Neogene–Quaternary volcanism pertinent to the recent stage of the Earth development is of a great scientific and practical importance. A possibility to employ three to four methods of isotopic dating of young volcanic rocks (K–Ar, Rb–Sr, U–Io, ^{14}C) creates the quantitative (numerical) base for stratigraphy, geological mapping, regional geodynamic reconstructions, and theory of volcanogenic ore formation.

Specific aspects of isotopic dating of young volcanics are well known [1]. In geological and petrological practice, the method elucidates the real duration of volcanic processes and active life of volcanoes. The isotope dates determine periods of magmatic activity and time of the latest eruptions of Quaternary volcanoes and, thus, are important for the forecast possible catastrophes.

The last aspect greatly stimulated in the last 7–10 years the isotopic-geochronological investigations of the Quaternary volcanoes in Japan, North America, and the Mediterranean belt, the Caucasus included.

Methodically, the isotope dating of young, especially of Neopleistocene–Holocene (younger than 1 Ma) volcanic rocks is the most complicated, especially the K–Ar dating, which can be universally used for the entire Quaternary period, from its beginning up to 20–30 ka, i.e., to the Holocene. The modern K–Ar dating of Quaternary volcanics is based on mass-spectrometer facilities to measure minor amounts (0.00*n*–0.0*n* ppb) of the radiogenic ^{40}Ar.

In this work, we present the results of geochronological investigations of young volcanism in the Lesser Caucasus exemplified by rocks of the Dzhavakheti neovolcanic region. The work is a part of our systematic study on isotopic geochronology and geochemistry of the young Caucasian magmatism. The Dzhavacheti region (upland) is a typical volcanic area of Lesser Cau-

casus in Georgia and Armenia, where Neogene–Quaternary volcanics of different chemical composition and age are widespread.

According to the common opinion [2], the intense Neogene–Quaternary magmatism of Caucasus is of the late collisional type by geodynamic nature. It commenced the middle Miocene, when the Caucasian segment of the Alpine folded belt was involved in the process of "rigid" collision between the Arabian and Eurasian continental plates, and lasted to the Holocene.

Zaridze [3], Milanovskii and Koronovskii [4], and other researchers distinguished three stages of Neogene–Quaternary magmatism in the Caucasus: (1) the late Miocene–early Pliocene, (2) late Pliocene, and (3) late Pliocene–Anthropogene stages of volcanic activity. The first stage is most fully recorded in the Lesser Caucasus, whereas magmatic rocks of the second stage are widespread throughout the Caucasian segment. Volcanism of the third (Quaternary) stage was of a local significance and resulted in formation of the big Elbrus, Kazbek, Aragats, and others stratovolcanoes.

On the base of isotopic-geochronological data recently obtained in the Greater Caucasus, time limits and occasionally durations of stages and phases of the magmatic activity have been determined. At the first stage (9.5–8.3 Ma) laccoliths of the Pyatigorsk area [5] were formed. According to results obtained by Arakelyants *et al.*, [6] and in later works [7], the second stage (about 4–2 Ma) resulted in origin of numerous hypabyssal granite intrusions (the El'dzhurtu, Tepli, Songutidon and other massifs) and ignimbrites of the Verkhnii and Nizhnii Chegem plateaus. Chernyshev *et al.* [8] published the first reliable K–Ar dates for Quaternary volcanics of the Kazbek area and Elbrus volcanic center. These and radiocarbon dates [9] showed that recent volcanism in the Greater Caucasus developed in three discrete phases during the last 450 ka. One more phase of explosive activity in the Elbrus area was also suggested. Thus, summarizing the available isotopic-geochronological information, we can distinguish three stages of magmatic activity in the Greater Caucasus, which took place in (1) the late Miocene (9.5–8.3 Ma), (2) Pliocene (4–2 Ma), and (3) Quaternary (900–0 ka).

The systematic isotopic-geochronological study of the Neogene–Quaternary volcanism of Lesser Caucasus has not been carried out until recently. Single age estimates were insufficient for geochronological scales and reconstructions. The progress in geochronology of young magmatism in the region was also decelerated by a considerable amount of material erupted during the last 10 m.y. and widespread here from the Adzhariya to Nagornyi Karabakh. The west–east succession of the Erushet-Arsiani, Dzhavacheti, Aragats, Gegam, Vardenis, Syunik and Araks volcanic zones are distinguishable in the region [10]. In the south, the volcanic province of the Lesser Caucasus is contiguous with the Kars and Ararat zones of similarly young magmatism. The products of Neogene–Quaternary magmatic activity in the Lesser Caucasus demonstrate a wide petrochemical spectrum from olivine basalts to rhyolites and obsidians. In this work, we discuss the results of isotopic-geochronological investigations of moderately acidic to intermediate volcanic rocks of the Dzhavakheti area, which are commonly united into the Goderdzi Formation [11].

BRIEF GEOLOGICAL REVIEW OF THE DZHAVAKHETI VOLCANIC AREA

The Dzhavakheti volcanic area (Fig. 1) is located in the central part of the Lesser Caucasus mountain system. It is bordered by the Gek-Tapa (Egnakhag), Shirak and Bazum ridges in the south, by upper courses of the Kura River in the west, and by the Trialeti Ridge of Paleogene volcanogenic–sedimentary deposits in the north. The eastern boundary of the young volcanic area coincides with the Tsalka–Gomere–Dmanisi–Stepanavan line, and three large flows of basaltic lava slope down from that line for tens of kilometers along the Mashavera, Khrami and Debed river valleys, being traceable up to the Khrami River mouth.

The Dzhavakheti Upland represents a medium to high mountainous land with the maximal peaks of about 3300 m high (the Didi-Abuli and Samsari mountains of the Samsari Ridge). The area represents a combination of mountainous plateaus and volcanic ridges with separate central volcanoes, extrusive domes, and intermountain depressions.

The basement of the Dzhavakheti area is composed of Cretaceous and Paleogene volcanogenic–sedimentary deposits, which are exposed in separate erosional windows in many places. In the east (the Khrami valley), there are exposures of the Khrami massif of Paleozoic granitoids. In the largest part of the region, the basement is overlain by Neogene–Quaternary volcanics, which are a few hundred meters thick.

According to modern views on geology of the region, the young volcanics of the Dzhavakheti area constitute the Goderdzi ($N_1^3 - N_2^1$), Akhalkalaki ($N_2^3 - Q_E$) and Kechut (Q_{II-III}) formations [12]. The formations ages correspond to time ranges of three stages of the Caucasian young volcanism, as established by Milanovskii and Koronovskii [4].

The Goderdzi Formation includes mainly intermediate and moderately acidic rocks. The Mio-Pliocene volcanism occurred in the vast territory from the southern fault of the Adzharia–Trialeti fold system in the north to the Lake Van in the south. Due to the general explosive character of volcanism, the Goderdzi Formation is composed of pyroclastic material and infrequent lavas. The classic Goderdzi sections are in the Erushet–Arsiani and Kars areas outside the Dzhavakheti Upland. Their basal beds representing by thin basic lava flows are overlain by thick pyroclastic unit of andesite–dacitic lava flows. The sections are crowned by andes-

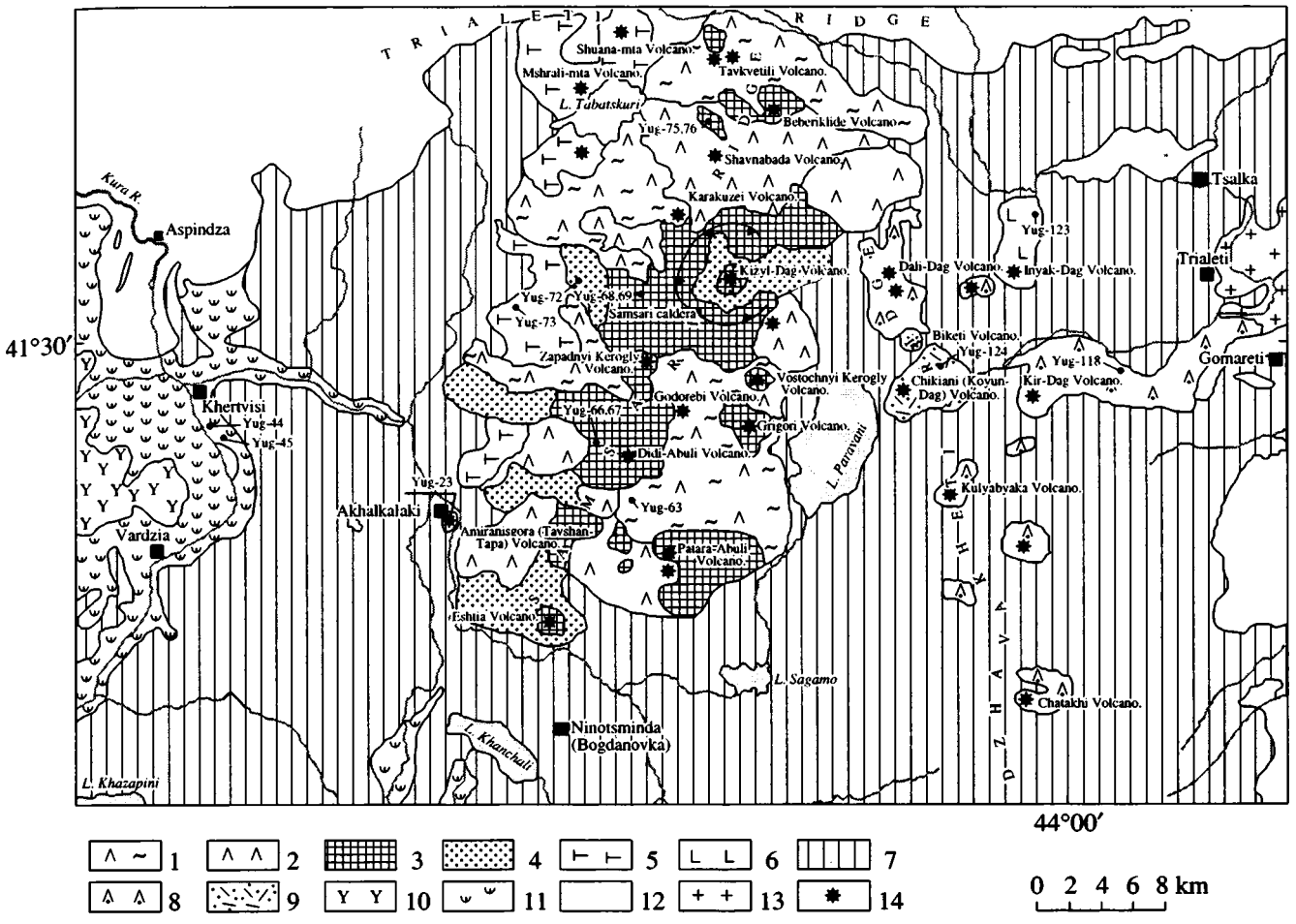

Fig. 1. Schematic geological map of the Dzhavakheti volcanic area (after Skhrtladze [11], Dzotsenidze and Kuloshvili [34], and Tutberidze [10]). (1–5) Quaternary volcanics of the Samsari Ridge: (1) hyalodacites (Phase III of volcanic activity), (2) pyroxene–plagioclase dacites (Phase II), (3) amphibole–plagioclase dacites (Phase II), (4) pyroclastics (Phase II), (5) two-pyroxene–amphibole–plagioclase dacites of the volcanoes of the "western range" of the Samsari Ridge (Phase I); (6) basaltic andesine of the Inyak-Dag Volcano (N_2); (7) basalts, basaltic andesites and andesites of the Akhalkalaki Formation (N_2); (8) dacites of the Dzhavakheti Ridge (N_2); (9) rhyolites, obsidians, perlites (N_2); (10) andesitic and dacitic lavas of the Goderdzi Formation (N_1); (11) pyroclastic rocks of the Goderdzi Formation (N_1); (12) Paleogene–Cretaceous volcanogenic–sedimentary deposits; (13) Paleozoic granitoids of the Khrami massif; (14) volcanoes.

itic and dacitic lavas. The Goderdzi Formation of the Erushet–Arsiani area is referred to the Mio-Pliocene based on stratigraphic data and floral remains [11].

In the Dzhavakheti region, pyroclastic rocks and andesitic lava flows of the Goderdzi Formation are exposed in the Kura and Paravani river valleys, where they are overlain by basalts of the Akhalkalaki Formation. Extrusive domes and volcanoes of the Samsari and Dzhavakheti ridges were also referred to the Goderdzi formation. These structures predominantly composed of dacites are confined to two large submeridional faults. Thus, the intermediate and moderately acidic rocks of the Dzhavakheti region are included into the Goderdzi Formation. It was supposed that dacitic lavas of the Samsari and Dzhavakheti ridges constitute upper parts of the formation, and that its lower pyroclastic part is not exposed here [11]. However, some geological peculiarities of the Dzhavakheti region evidence against the idea to unite all moderately acidic volcanics into the single formation. For instance, the dacitic lavas underlie basalts of the Akhalkalaki Formation in the Kura and Paravani river valleys and overlie them near the Korkhi-Ordzha Village. The late Miocene age of the Samsari Ridge volcanoes was doubted by Milanovskii and Koronovskii [4] and by other researchers earlier. Up to recently, the age problem of all moderately acidic rocks of the Dzhavakheti region was unsolved because of their different positioning in the Goderdzi Formation sections.

The Goderdzi pyroclastics are overlain by olivine basalts and basaltic andesites of the Akhalkalaki Formation. Lava flows of the latter leveled the ancient relief and formed the large Akhalkalaki, Tsalka, Gamer and other lava plateaus. Sequences of up to 20 lava

flows of the Alkhalkalaki Formation, which are locally up to 300 m thick in total, can be observed in ravines of the Kura, Paravani and other large rivers. Fauna-bearing lacustrine deposits in the middle part of the Akhalkalaki Formation (near the town of Akhalkalaki) reliably indicate the late Pliocene–Eopleistocene age of the basalts [13]. The youngest structures of the Akhalkalaki Formation are numerous cinder cones and andesitic lava flows, which occur as small hills on the basaltic plateaus [12].

The youngest deposits of the Dzhavakheti Upland are pyroxene dacites and hyalodacites of the Kechut Formation. Lava flows of this composition, which lope down from parasitic centers of the Samsari Ridge volcanoes, are neither long nor thick. Characteristic "stone chaos" on the flow surfaces and good preservation of the eruption centers suggest the young age of the Kechut Formation. The middle–late Neopleistocene age estimated for the formation [12] has not been proved however by either paleontological evidence or geochronological data.

Thus, the geological structure of the Dzhavakheti Upland is indicative of three stages of the young volcanic activity separated by considerable hiatuses.

FORMER GEOCHRONOLOGICAL DATA

Remains of a prehistoric man found in the Dmanisi area in the last decade, stimulated works on age determinations for the Akhalkalaki basalts. The time span of two plateau basalt sequences that was determined by the K–Ar and $^{40}Ar/^{39}Ar$ methods [14] is 2.9–1.8 Ma. This is consistent with the fauna-based age of the Akhalkalaki Formation. The other $^{40}Ar/^{39}Ar$ date (3.60 ± 0.06 Ma) was obtained for plagioclase phenocrysts from the lowest basalt flows of the Toki section in the Paravani valley [15]. The isotopic-geochronological investigations of volcanic rocks from the Aragats area [16] yielded the K–Ar date of 2.5 ± 0.2 Ma for the plateau basalts of the Akhutyan canyon in Armenia, which are analogues of the Akhalkalaki Formation rocks in the Dzhavakheti Upland. Accordingly, the Akhalkalaki Formation age should be within the range of 3.6–1.8 Ma (middle–late Pliocene). Paleomagnetic study of the Akhalkalaki basalts [17] revealed regular reversals of magnetic polarity in their sequence.

The age of the Akhalkalaki plateau basalts is the minimum age limit for the Goderdzi pyroclastics and lavas in the Kura and Paravani valleys. In the 1960s–1980s, two K–Ar dates for the Goderdzi pyroclastics were obtained in the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry (IGEM), one of 8.0 ± 1.5 Ma (ignimbrite, the Vardzia area, collection of E.K. Ustiev) and the other one of 6.6 $\pm$ 0.4 Ma (rhyolitic tuff, the Vardzia area, collection of V.N. Volkov). The K–Ar dating of the Goderdzi pyroclastic rocks that was done at the same time in the Geological Institute, Academy of Sciences of Georgia (GIN), approximately indicated their Mio-Pliocene age [18]. Stankevich [19] published paleomagnetic data for pyroclastics of the Goderdzi Formation in the Khertvisi Village area and for volcanoes Samsari, West Kerogly, and Ivan-Tepe of the Samsari Ridge, which, as mentioned above, are assigned to the upper part of the Goderdzi Formation. The Samsari Ridge volcanoes and the Late Quaternary volcanoes of the Kazbek area have similar paleomagnetic parameters, which essentially differ from those of the Goderdzi pyroclastic rocks and cast doubts on their coeval origin. The only obsidian dome of the Chikiani (Koyun-Dag) Volcano in southern Georgia, which was also referred to the Goderdzi Formation, was dated at 2.2–3.0 Ma by the fission track method [20]. This structure is therefore closer in age to the Akhalkalaki Formation rather than to the Goderdzi pyroclastics. In general, the volcanics referred to the Goderdzi Formation have not been systematically studied, and their age remains uncertain.

Geochronological investigations of the Kechut Formation volcanics have never been carried out. A single date of 0.37 ± 0.15 Ma (middle–late Neopleistocene) was obtained for hyalodacites of the Tavkvetili Volcano in GIN (Rubinshtein *et al.*, 1972).

PETROGRAPHY AND CHEMICAL COMPOSITION OF THE ROCKS STUDIED

In 1999–2000, the joint expedition of IGEM and GIN collected representative samples of volcanic rocks of different age and composition from different parts of the Dzhavakheti volcanic area. Described below are predominantly the samples of volcanics of the Mio-Pliocene Goderdzi Formation, lavas of the Samsari and Dzhavakheti ridges, volcanic rocks of the Kura valley, and some extrusive structures of the region included. Samples of the youngest Kechut Formation were also studied. Localities of sampling sites are shown in the schematic geological map of the Dzhavakheti Upland (Fig. 1) and in Table 1.

The Dzhavakheti volcanics can be divided into several petrographic groups.

Biotite–amphibole–plagioclase dacites of the Amiranisgora (Tavshan-Tapa) Volcano (samples Yug-2 and Yug-3). Phenocrysts constitute in total 10–15% of the rock volumes. Plagioclase of andesine composition occurs as occasionally zoned prismatic and xenomorphic crystals. Some crystals are partly resorbed. Plagioclase phenocrysts vary in size from 0.1 to 8 mm. Elongate, less frequently isometric crystals of basaltic hornblende are 0.2 to 1.2 mm in size and larger (up to 2 mm). Their idiomorphic hexagonal forms are rare, and absence of secondary opacite indicates hypabyssal origin of the rocks. Scarce biotite (its greatest abundance was recorded in the contact zone of extrusions) occurs as single elongate crystals 0.2 to 0.4 mm in size. The groundmass represents to 85–90% of the rocks. It is commonly hyalopilitic and locally fluidal, when ori-

Table 1. Brief characteristics of the studied geological samples from the Dzhavakheti volcanic upland

Sample number	Rock	Geological object	Sampling site
Yug-2	Biotite–amphibole dacite (pink)	Amiranisgora (Tavshan-Tapa) Volcano	SE-E slope of Amiranisgora Volcano, altitude 1750 m
Yug-3	Biotite–amphibole dacite (gray)	"	"
Yug-44	Two-pyroxene dacite	Intraformational flow in the pyroclastics of the Goderdzi Fm.	2 km to the south of Khertvisi Village along the Khertvisi–Vardzia road
Yug-45	"	"	3 km to the south of Khertvisi Village along the Khertvisi–Vardzia road
Yug-63	Hyalodacite	Kechut Fm., lava flow of Goderebi Volcano	S slope of Didi-Abuli Volcano, Abuli–Vladimirovka road, 200 m below the pass between Didi-Abuli and Patara-Abuli volcanoes
Yug-66	Amphibole dacite (gray–pink)	Didi-Abuli Volcano	W-NW slope of Didi-Abuli Volcano, 2 km to the top; 5 km to NE of Buzaveti Village at altitude of 2600 m
Yug-67	Amphibole dacite (pink)	"	"
Yug-68	Amphibole dacite (gray)	Samsarid Volcano	NW slope of Samsari Volcano, locality on Samsari Volcano slope 4 km to NW of the top, altitude 2300 m
Yug-69	Amphibole dacite (pink)	"	"
Yug-72	Amphibole dacite (gray)	Pyroclastics of Samsari Volcano	Cave-town below Patara-Samsari Village, quarry
Yug-73	Two-pyroxene–amphibole dacite	Babakhngo Mt.	W slope of Babakhngo Volcano, Didi-Samsari–Merenia road, 1 km of Merenia Village
Yug-75	Amphibole dacite	Shanabada Volcano	N slope of Shavnabada Volcano, Tabatskuri-Tsalka Mt. road, 100 m of Krasnyi Arbovshchik yaila
Yug-76	"	"	Altitude 2172.3 m on N slope of Shavnabada Volcano, pass between Tabatskuri and Tsalka Mt.
Yug-118	Aphyric dacite	Kir-Dag Volcano lava flow	Pass between Tsalka Mt.–Gomareti Mt. across the ridge separating Tsalka and Gomareti plateaus
Yug-123	Aphyric basaltic andesite	Inyak-Dag Volcano lava flow	Tikilis area, 1 km of Tsalka Mt.–Lake Paravani to the south
Yug-124	Obsidian	Chikiani (Koyun-Dag) Volcano	NE slope of Chikiani Volcano, obsidian quarry

Note: Rocks are named based on their chemical composition.

ented microlites flow around the phenocrysts. Set in the glassy matrix are andesine microlites, small crystals of hornblende, ore minerals, and cristobalite. Traces of superimposed processes can be observed. Samples Yug-2 and Yug-3 differing the groundmass color (pink and gray respectively).

Two-pyroxene–plagioclase dacites from the intraformational flows of the Kura valley (samples Yug-44 and Yug-45). Phenocrysts of plagioclase, monoclinic and rhombic pyroxenes constitute more than 50% of the rock volumes. Idiomorphic prismatic and infrequent xenomorphic, partially resorbed crystals of plagioclase (andesine) vary in size from 0.1 to 1 mm, less frequently to 2.5 mm. The total plagioclase content is 50%. Plagioclase phenocrysts in Sample Yug-45 contain abundant gas–fluid and melt inclusions. Idiomorphic clinopyroxene crystals (up to 5% of the rock volume) range in size from 0.1 to 0.9 mm, less frequently to 1.5 mm. Small (0.1–0.5 mm) orthopyroxene crystals (less than 1%) show traces of alteration along transverse parting. Predominantly hyalopilitic groundmass is composed of plagioclase microlites and partly recrystallized glass with dusty inclusions of ore minerals. Amygdules in Sample Yug-45 are filled with aggregate of anisotropic matter.

Hyalodacite of the Kechut Formation from the lava flows of the Godorebi Volcano (Sample Yug-63). The rock is black, with conchoidal fracturing and porphyritic texture. There are phenocrysts of plagioclase and hornblende. Rare idiomorphic plagioclase crystals 0.3–0.4 mm in size constitute not more than 3 or 4% of the rock. Single hornblende crystals are entirely replaced by opacite. The hyalopilitic groundmass is composed

Table 2. Chemical composition of the studied volcanics of the Dzhavakheti area

Sample no.	SiO_2, %	Al_2O_3, %	Fe_2O_3, %	MnO, %	TiO_2, %	Na_2O, %	K_2O, %	CaO, %	MgO, %	P_2O_5, %	Rb, ppm	Sr, ppm
Yug-2	72.25	14.26	2.85	0.045	0.33	3.50	2.59	3.29	0.78	0.118	68	538
Yug-3	71.18	14.62	3.12	0.046	0.37	3.59	2.47	3.58	0.89	0.135	64	570
Yug-44	67.33	15.24	4.72	0.055	0.60	3.56	2.11	5.09	1.03	0.266	43	624
Yug-45	65.48	14.54	5.47	0.065	0.76	3.63	2.56	5.56	1.55	0.387	52	774
Yug-63	64.71	16.09	6.04	0.082	0.85	3.98	2.21	4.64	1.12	0.279	41	609
Yug-66	71.08	14.50	3.25	0.045	0.40	3.52	2.13	4.26	0.69	0.121	52	559
Yug-67	68.27	14.72	3.43	0.049	0.40	3.51	2.03	6.58	0.89	0.126	48	578
Yug-68	67.07	15.71	3.89	0.053	0.50	3.65	1.92	6.36	0.72	0.126	45	610
Yug-69	67.91	15.65	3.98	0.057	0.52	3.78	1.90	5.30	0.78	0.132	43	647
Yug-72	70.31	14.43	3.83	0.054	0.50	3.60	2.25	4.13	0.74	0.159	51	590
Yug-73	66.98	15.74	4.68	0.062	0.55	3.68	1.66	5.40	1.12	0.140	38	561
Yug-75	71.75	14.10	2.84	0.043	0.40	3.76	2.19	4.19	0.62	0.115	54	613
Yug-76	66.67	15.79	4.62	0.067	0.56	3.93	1.78	5.08	1.31	0.195	32	611
Yug-118	69.32	15.76	3.64	0.050	0.55	3.63	2.37	3.92	0.65	0.125	68	615
Yug-123	56.48	17.69	7.84	0.123	1.08	3.83	1.82	7.68	3.06	0.406	27	604
Yug-124	77.04	12.49	1.28	0.063	0.14	3.52	4.52	0.75	0.17	0.016	124	90

Note: Rocks are analyzed by A.I. Yakushev and T.M. Marchenko, IGEM, the X-ray fluorescent spectroscopy, spectrometer Philips PW 2400. Analytical results are recalculated for 100% tot.

predominantly of volcanic glass with dusty inclusions of ore mineral and oriented plagioclase laths.

Amphibole–plagioclase dacites of the Samsari Ridge volcanoes (Didi-Abuli, Samsari, Shavnabada) (samples Yug-66, Yug-67, Yug-68, Yug-69, Yug-72, Yug-75, and Yug-76). The summary content of plagioclase and basaltic hornblende phenocrysts ranges from 30 to 50%. Prismatic or resorbed xenomorphic plagioclase crystals occasionally show indistinct zoning. In phenocrysts, their size is between 0.2 and 1.2 mm. The content of plagioclase phenocrysts in dacite samples ranges from 10–15 to 45%. Hornblende phenocrysts 0.2–1.0 mm in size are partly or entirely replaced by opacite. Only central parts of hornblende crystals usually remain unchanged. The hornblende content is from 5 to 10–15%. Groundmass usually fluid is sometimes hyalopilitic and pilotaxitic. It is composed of volcanic glass, plagioclase laths, magnetite grains, and small occasional hornblende crystals. Rocks are lacking traces of secondary alterations.

Two-pyroxene–amphibole–plagioclase dacite of the Babakhngo Mountain (Sample Yug-73). The rock of porphyritic texture is gray and vesicular in places. It contains phenocrysts of plagioclase, hornblende, clinopyroxene, and orthopyroxene. The plagioclase phenocrysts are represented by indistinctly zoned idiomorphic prismatic crystals varying in size from 0.15 to 1.0 mm. The plagioclase of andesine composition constitutes to 25% of the rock. The basaltic hornblende phenocrysts (about 5% of the rock) are replaced partly or entirely by opacite. They are of prismatic habit and range in size from 0.1 to 2.0 mm. Isometric and partially resorbed clinopyroxene crystals (1–2% of the rock) are 0.1–0.2 mm in size and show reaction rims at the contacts with groundmass. Single prismatic crystals of orthopyroxene are up to 0.3 mm in size. The microlitic groundmass is composed of plagioclase microlites with subordinate amount of volcanic glass.

Weakly porphyritic dacite of the Kir-Dag Volcano (Dzhavakheti Ridge) (Sample Yug-118). This microcrystalline rock is pink–brown in color and porphyritic in texture. The content of phenocrysts do not exceed 10%. The phenocrysts are represented by idiomorphic and less frequent xenomorphic crystals of plagioclase (andesine) 0.3–0.7 mm in size and by single grains of hornblende entirely replaced by opacite. Plagioclase crystals are indistinctly zoned. The trachytic groundmass is composed of subparallel plagioclase laths and elongated magnetite crystals of, which completely substitute hornblende.

Basaltic andesite from the lava flow of the Inyak-Dag Volcano (Sample Yug-123). The rock is aphyric and microdoleritic in texture. It is composed of randomly oriented laths of fresh plagioclase and small isometric clinopyroxene crystals in between.

Obsidian of the Chikiani Volcano (Sample Yug-124). The rock is black, showing a weak glass devitrification in places.

Thus, the moderately acidic volcanics of the Dzhavakheti Upland can be classified into six petrographic types of dacites according to the phenocryst paragene-

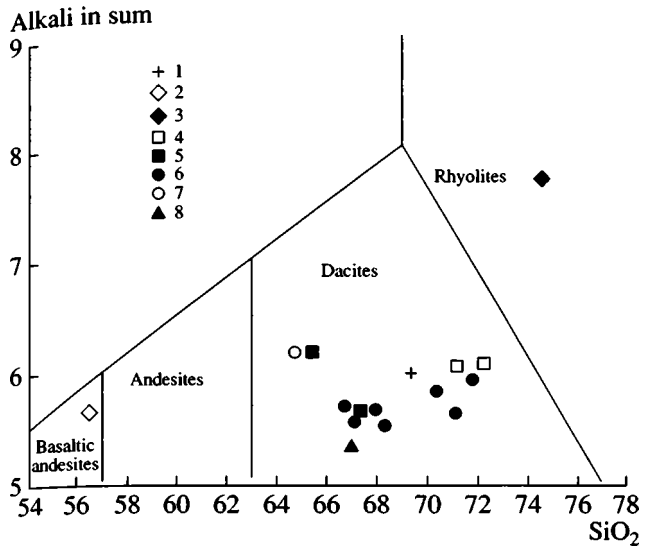

Fig. 2. The (Na_2O+K_2O)–SiO_2 classification diagram for the studied rocks of the Dzhavakheti area (fields for rock types are those recommended by the IUGS Subcommission on Systematics of Igneous Rocks [21]). Symbols for Figs. 2–5 and 7: (1) Kir-Dag (Dzhavakheti Ridge), (2) Inyak-Dag, (3) Chikiani, and (4) Amiranisgora volcanoes; (5) dacites, upper courses of the Kura River; (6) volcanoes of the Samsari Ridge; (7) hyalodacites of the Samsari Ridge; (8) Babakhngo Volcano (the western range of the Samsari Ridge).

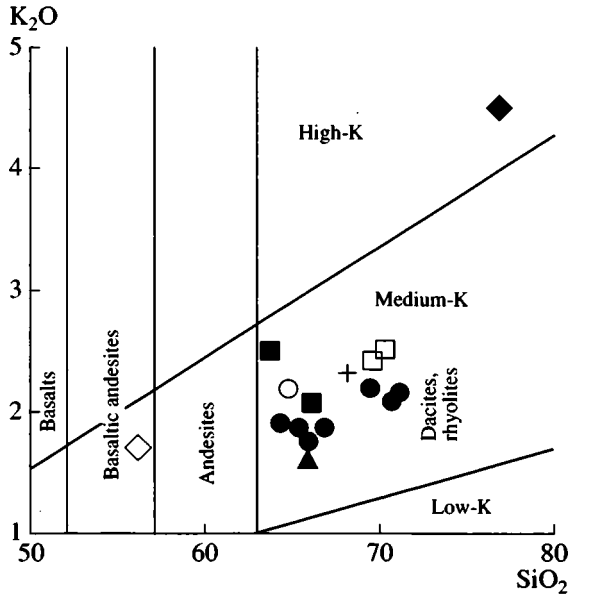

Fig. 3. The K_2O–SiO_2 classification diagram [21] for the studied rocks of the Dzhavakheti area (symbols as in Fig. 2).

sis These are biotite–amphibole–plagioclase, two-pyroxene–plagioclase, amphibole–plagioclase, and two-pyroxene–amphibole–plagioclase dacites, hyalodacites, and weakly porphyritic dacites of the Kir-Dag Volcano. The group does not include the Chikiani obsidians and Inyak-Dag basaltic andesites, which were also referred formerly to the Goderdzi Formation (for example, [11]).

In chemical composition, the majority of studied volcanics of the Dzhavakheti area corresponds to dacites with silica content varying from 64 to 72% (Table 2, Fig. 2). The exceptions are the obsidian of the Chikiani Volcano (Yug-124) and the rock of the lava flows from the Inyak-Dag Volcano (Yug-123), which correspond in composition to rhyolite and basaltic andesite respectively (Fig. 2). All the volcanics, except for the basaltic andesite Yug-123, have a high percentage of normative quartz (20–35%).

Among samples from the Dzhavakheti volcanic area, hyalodacite of the Kechut Formation (Yug-63) and rocks of intraformational flows in pyroclastic units of the Kura valley (Yug-44 and Yug-45) have the lowest silica content. The rocks of the Amiranisgora extrusive are most acidic. Volcanics of the Samsari Ridge (Fig. 2) are intermediate between them.

In SiO_2–K_2O classification diagram [21], data points of the Dzhavakheti volcanics studied are mostly concentrated in the field of rocks with moderate potassium content (Fig. 3). The only exception is the highly potassium Chikiani obsidian (Yug-124).

The studied rocks of the Dzhavakheti Upland (all the dacite varieties and basaltic andesite of the Inyak-Dag Volcano) belong to the calc–alkaline series (Fig. 4), whereas the Chikiani Volcano obsidian (Yug-124) belongs to the K–Na subalkaline series. This emphasizes a specific position of this magmatic rock among other intermediate and acidic volcanics of the Dzhavakheti Upland. Figs. 3 and 5 show variations of main oxides concentrations versus SiO_2 content in the rocks. It is clearly seen that contents of CaO, Fe_2O_3, P_2O_5, and TiO_2 diminish as the silica content increases. These trends are less pronounced for Al_2O_3, MgO, and MnO. It is difficult to specify any regular trend of K_2O and especially Na_2O concentrations in the studied volcanics.

In connection with the Rb–Sr method applied, the Rb and Sr concentrations in the Dzhavakheti rocks were analyzed (Table 2). It is natural that the Rb content is low in the Inyak-Dag basaltic andesite (Yug-123) and high in the Chikiani obsidian (Yug-124). The Rb concentrations in Dzhavakheti dacites range from 30 to 70 ppm.

Virtually all the studied Dzhavakheti dacites show the high Sr content, that is 1.5–2 times higher in average than the clarke values for intermediate and acidic rocks, whereas Ca-concentrations are usual for rocks of this type. It is remarkable that correlation between concentrations of these geochemically allied elements is absent (Fig. 6).

The Rb–Sr versus SiO_2 diagrams for the Dzhavakheti rocks are shown in Fig. 7. The Rb concentration

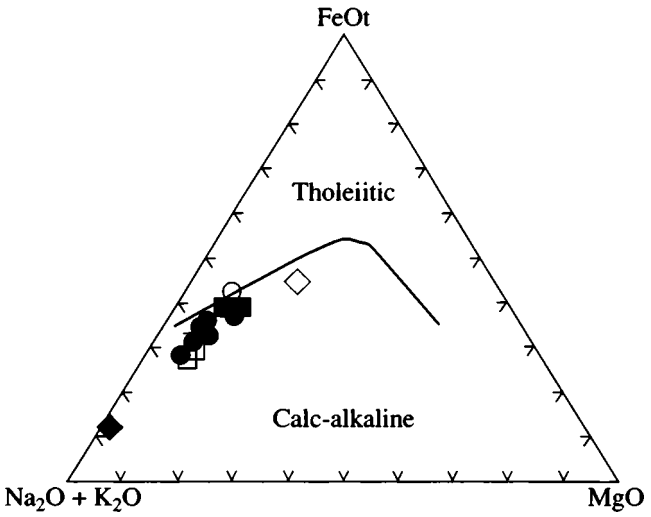

Fig. 4. The AFM diagram for the studied, moderately acidic volcanics of the Dzhavakheti area (symbols as in Fig. 2).

generally grows with the increase of silica content, whereas the Sr content is not silica-dependent.

The results obtained allow us to divide the Dzhavakheti volcanics referred previously to the Goderdzi Formation into three groups of rocks according to their petrographic and petrochemical features. The overwhelming majority of the volcanics is represented by calc-alkaline dacites, which build up the volcanoes of the Samsari and Dzhavakheti ridges and lava flows in the Kura valley. The Chikiani Volcano is composed of subalkaline acidic rocks. Basaltic andesites of the Inyak-Dag Volcano are similar in chemical composition to the plateau basalts of the Akhalkalaki Formation.

METHODS OF ISOTOPIC-GEOCHRONOLOGICAL ANALYSIS

We used both the K–Ar and Rb–Sr isotopic methods to study the volcanics of the Dzhavakheti Upland.

The largest part of geochronological dates is obtained by the K–Ar method. The K–Ar dating of young volcanics is based on measurements of ultraminor quantities of radiogenic argon, which usually constitutes an insignificant portion of argon in the samples. The Dzhavakheti volcanics were studied using a special technique developed in IGEM [22] and the highly sensitive mass-spectrometer constructed at that institute based on device MI-1201 IG of the Sumy NPO SELMI (Ukraine). The radiogenic argon was measured by the isotope dilution technique with the ^{38}Ar as a spike. The analytical procedure and precision allowed dating the samples weightings not more than 50 mg. The measurements were controlled by the regular analyses of standard samples: biotite–muscovite "Bern-4," basalt "1/76 Asia", muscovite "P-207," and atmospheric argon. The K concentration was measured by the flame spectrophotometry method.

The previous study of the K–Ar isotopic systematics for phenocrysts and groundmass of volcanics from the Greater Caucasus showed that the latter substance was best suitable for dating because the argon-saturated phenocrysts (especially with low K content) yielded overestimated ages [8]. The same effect is known for the Quaternary lavas of Japan, Italy, and North America, as is reported in foreign works [23]. Therefore, the groundmass freed of the phenocrysts was used for the K–Ar dating of the young Dzhavakheti volcanics. The whole-rock method was applied to samples Yug-44 and Yug-45, whose Miocene age was proved by preliminary studies. In addition, we studied the K–Ar isotopic systematics for phenocrysts from some samples of the Samsari Ridge volcanics. Prior to the isotope analysis, the dacite sample Yug-45 was treated in a weak HNO_3 solution to remove carbonate from amygdules.

The traditional ion-exchange chromatographic technique was used to extract Rb and Sr for isotopic analysis. Their concentrations were measured by the isotope dilution technique with the mixed spike $^{85}Rb + {}^{84}Sr$. The isotopic analyses were made using the mass-spectrometer Micromass Sector 54. The Sr isotope ratio for the standard sample SRM-987, which was regularly measured during the experimental runs, was 0.710248 ± 0.000011. Errors for $^{87}Sr/^{86}Sr$ and $^{87}Rb/^{86}Sr$ ratios in the samples studied are not greater than 0.003 and 1.7% respectively.

The Nd isotope composition, Sm and Nd concentrations were measured with the mass-spectrometer Micromass Sector 54. The $^{143}Nd/^{144}Nd$ ratio in the Nd standard of La Jolla, which was analyzed during experiments, corresponds to 0.511799 ± 0.000010. Errors for the $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios in the samples studied did not exceed 0.018 and 0.09%, respectively.

Geochronological calculations are based on the decay constants recommended by the International Subcommission on Geochronology [24].

RESULTS OF ISOTOPIC-GEOCHRONOLOGICAL INVESTIGATIONS AND DISCUSSION

The K–Ar isotopic ages are determined for 22 samples of moderately acidic volcanics from the Dzhavakheti area. The results are shown in Table 3.

Dacitic Lavas of the Upper Course of the Kura River

The K–Ar whole-rock analysis was applied to date the intraformational two-pyroxene–plagioclase dacite flows in the pyroclastic unit of the Goderdzi Formation (Yug-44 and Yug-45, the Kura valley). The obtained dates (7.6 and 7.3 Ma) are concordant within the analytical error limits and correspond to the supposed age (late Miocene) of the enclosing Goderdzi pyroclastics. It should be noted that the dates are close to the K–Ar ages (8.0 ± 1.5 and 6.6 ± 0.4 Ma respectively) obtained previously in the laboratory of IGEM for ignimbrites of the Goderdzi Formation (collections of E.K. Ustiev and

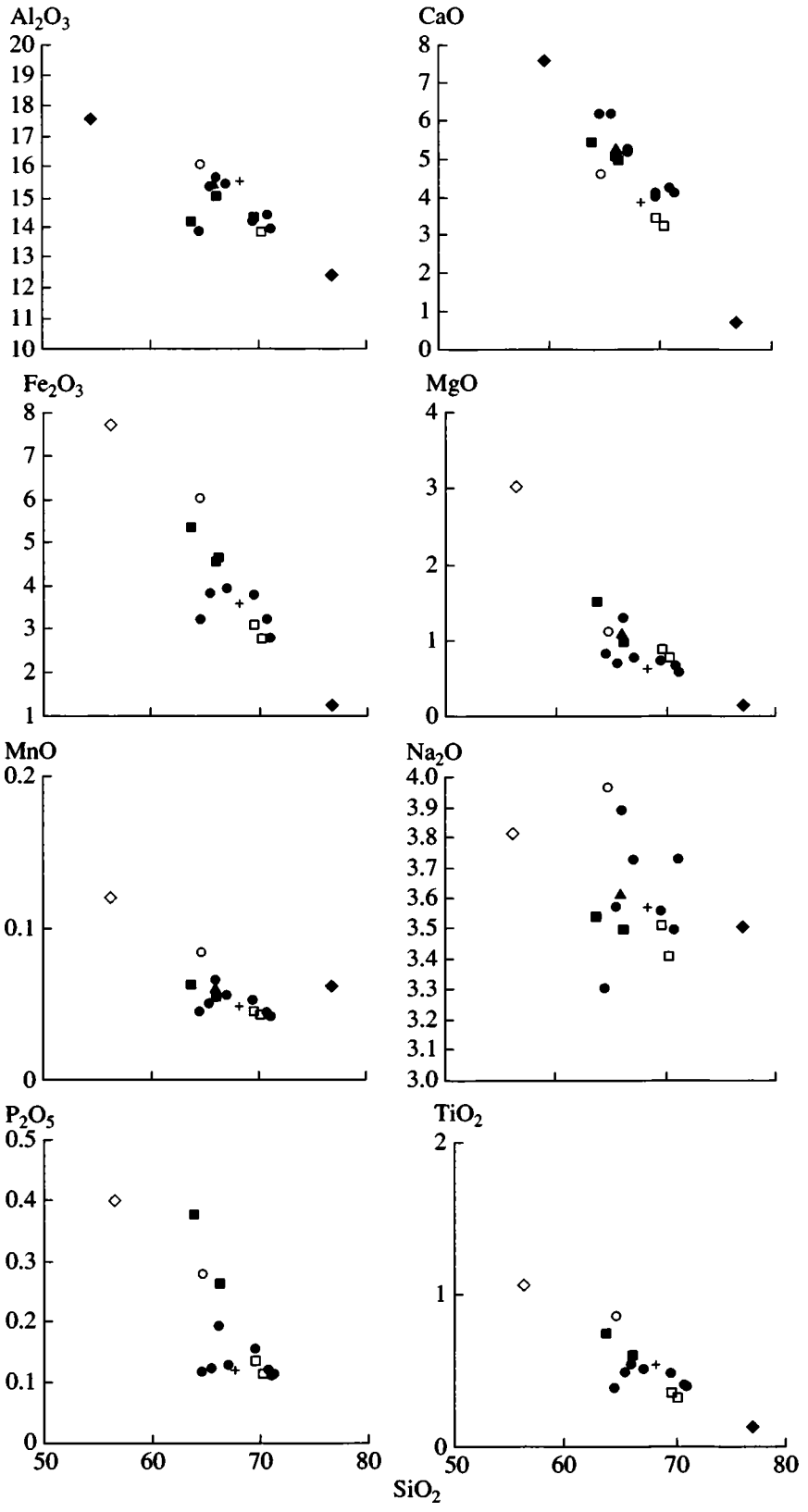

Fig. 5. Diagram of main oxides versus SiO_2 in volcanic rocks of the Dzhavakheti area (symbols as in Fig. 2).

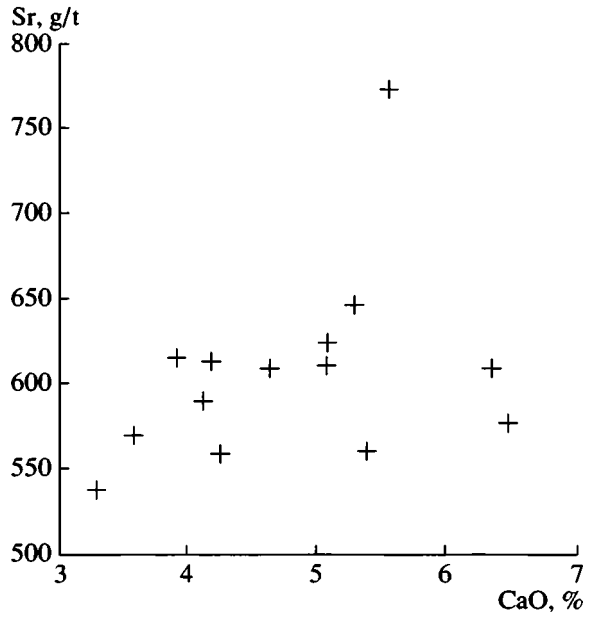

Fig. 6. The CaO–Sr correlation diagram for the studied dacites of the Dzhavakheti area.

V.N Volkov, the Kura valley). As mentioned above, the late Miocene age of the Goderdzi pyroclastics was based on fossil flora found in the tuffs of the Erusheti-Arsiani Upland joining to the Dzhavakheti area in the west. In this upland, the pyroclastics are widespread and traces of later magmatic processes are lacking. It is interesting that the K–Ar dating of teschenites from the Adzharia-Trialeti zone bordering the Dzhavakheti area in the north [25] yielded a close date of 7 ± 1 Ma. Thus, to the west of the Kura valley of southern Georgia, there was obviously a powerful impulse of predominantly explosive volcanism with subordinate eruptions of moderately acidic and even basic lavas (thin basaltic flows occur at the base of the pyroclastic unit in the Goderdzi Pass area) in the initial Late Miocene. To the east of the Kura valley (in the Dzhavakheti Upland), pyroclastic rocks are widespread along the Paravani valley up to the Toki Village area, where they underlie the basalts of the Akhalkalaki Formation. The established age (about 7.5 Ma) of the intraformational dacitic flows, which is likely close to that of the entire pyroclastic unit, the commencement time of the Late Cenozoic magmatic activity in southern Georgia after the considerable time gap since the Oligocene time.

The Amiranisgora and Chikiani Extrusive Domes

According to present views on geology of the Dzhavakheti Upland, dacitic lavas of the Samsari volcanoes and some extrusive structures composed of moderately acidic and acidic rocks are constituents, as mentioned above, of the Goderdzi Formation [11, 12]. These views are inconsistent with the K–Ar dates obtained for the Amiranisgora and Chikiani) extrusive structures com-

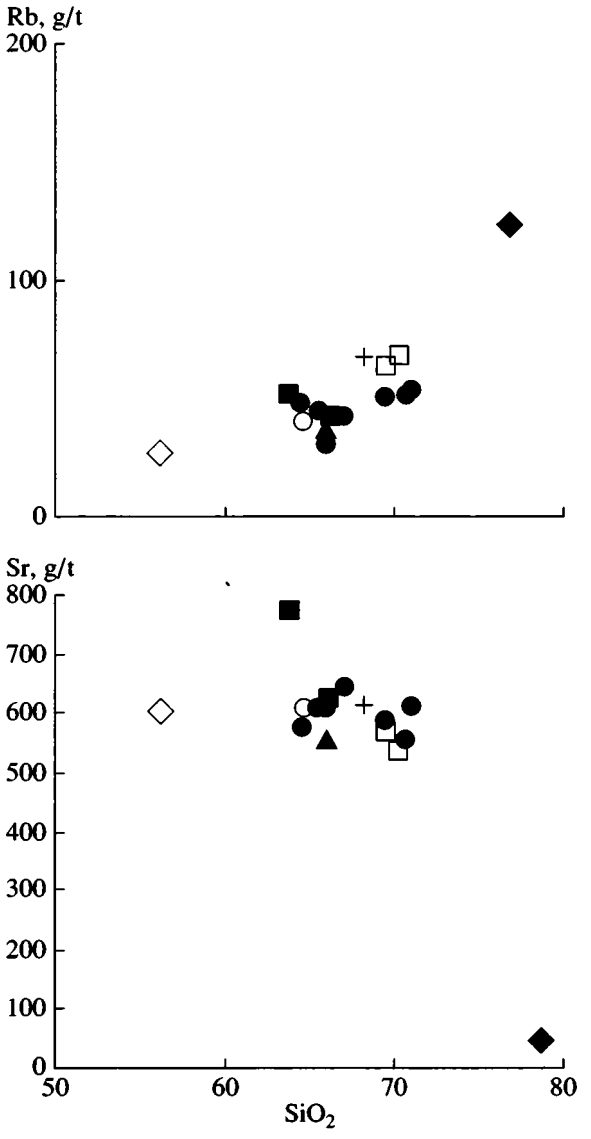

Fig. 7. Rb and Sr versus SiO_2 diagrams for the studied dacites of the Dzhavakheti area (symbols as in Fig. 2).

posed of the most acidic volcanic rocks of the Dzhavakheti Upland.

The date of 2.7 ± 0.15 Ma was obtained for obsidian of the Chikiani (Koyun-Dag) Mountain (Yug-124, Table 3). The K–Ar date of 2.85 ± 0.2 Ma (Table 3, at the bottom) that is close within the error limits to the above date was previously obtained for the Chikiani rhyolites (collection of Volkov) by one of us in IGEM [26]. The fission track age of the Chikiani obsidian is 3.0–2.2 Ma [20]. Consequently, the recently obtained and previous dates obviously indicate the late Pliocene age (2.85–2.7 Ma) of the Chikiani rhyolite–perlite–obsidian extrusion. The available geochronological data on the Akhalkalaki Formation [27] suggest that the Chikiani Volcano is concurrent in age to the plateau basalts. On the contrary, there is a considerable time

Table 3. K-Ar isotopic ages of the samples from the Dzhavakheti volcanic area

Sample no.	Rock	Sample	K, % ±σ	$^{40}Ar_{rad}$, ppb ±σ	$^{40}Ar_{atm}$, %	Age, Ma ±1.6σ
Yug-2	Dacite	Biotite	5.35 ± 0.05	1.77 ± 0.06	58.1	4.8 ± 0.3
		Plagioclase	0.30 ± 0.015	not detected	100	–
		Groundmass	2.33 ± 0.03	0.44 ± 0.03	93.9	2.7 ± 0.3
Yug-3	"	Biotite	4.76 ± 0.07	1.39 ± 0.06	74.6	4.2 ± 0.3
		Groundmass	2.14 ± 0.03	0.48 ± 0.02	78.6	3.2 ± 0.2
Yug-44	"	Whole-rock	1.87 ± 0.03	0.99 ± 0.04	16.3	7.6 ± 0.4
Yug-45	"	"	2.35 ± 0.03	1.19 ± 0.05	15.8	7.3 ± 0.4
Yug-63	"	Groundmass	1.89 ± 0.02	0.003 ± 0.003	99.7	0.03 ± 0.03
Yug-66	"	"	1.90 ± 0.02	0.037 ± 0.004	97.5	0.28 ± 0.06
Yug-67	"	Plagioclase	0.45 ± 0.015	0.12 ± 0.01	85.5	3.7 ± 0.4
		Groundmass	1.71 ± 0.02	0.029 ± 0.004	98.4	0.24 ± 0.06
Yug-68	"	"	1.68 ± 0.02	0.021 ± 0.004	99.0	0.18 ± 0.07
Yug-69	"	"	1.74 ± 0.02	0.029 ± 0.005	99.0	0.24 ± 0.07
Yug-72	"	"	1.70 ± 0.02	0.023 ± 0.005	95.8	0.20 ± 0.04
Yug-73	"	"	1.68 ± 0.02	0.089 ± 0.009	91.4	0.76 ± 0.08
		Plagioclase	0.21 ± 0.015	0.12 ± 0.01	61.0	8.2 ± 0.7
Yug-75	"	Groundmass	1.92 ± 0.02	0.039 ± 0.004	97.1	0.29 ± 0.06
Yug-76	"	Plagioclase	0.28 ± 0.015	0.024 ± 0.006	93.0	1.2 ± 0.3
		Groundmass	1.66 ± 0.02	0.037 ± 0.004	98.1	0.32 ± 0.07
Yug-118	"	"	2.09 ± 0.02	0.33 ± 0.01	94.8	2.25 ± 0.10
Yug-123	Basaltic andesite	"	1.44 ± 0.02	0.21 ± 0.01	92.1	2.05 ± 0.10
Yug-124	Obsidian	Glass	3.90 ± 0.03	0.73 ± 0.03	10.5	2.7 ± 0.15
1549-1*	Rhyolite	Whole-rock	3.94 ± 0.04	0.78 ± 0.04	29.7	2.85 ± 0.20

* Sample of V.N. Volkov's collection.

gap (about 4.5 m.y.) between the formation of the Goderdzi pyroclastic rocks and the Chikiani extrusion. In addition, the Chikiani mountain rocks essentially differ in chemical composition from other Goderdzi dacitic lavas and show affinity with subalkaline K–Na series (Table 2). All these facts suggest that rhyolite–obsidian volcanism of the Dzhavakheti Upland corresponds to an independent Upper Pliocene volcanic formation.

The biotite–amphibole–plagioclase dacites of the Amiranisgora (Tavshan-Tapa) Mountain near the Akhalkalaki area are most acidic among dacites of the Dzhavakheti Upland (Fig. 2, Table 2). Their pink and gray varieties are undoubtedly coeval: the gradual color changes and the absence of intrusive contacts have been described by many authors (e.g., [11]) and observed by us. As dacites are composed of several appropriate mineral phases, they are the most suitable rocks for isotope dating. Therefore, we used both the K–Ar and isochron Rb–Sr methods to determine the age of Amiranisgora dacites. The results obtained faile however to date unambiguously the Amiranisgora extrusion time. The K–Ar ages for groundmass of two dacite varieties appeared to be very close: the pink dacite (Yug-2) is 2.7 ± 0.3 Ma old and the gray dacite (Yug-3) is 3.2 ± 0.2 Ma old. These dates approximate the age of the Chikiani rhyolite–perlite–obsidian extrusion. The biotite fraction of dacites yielded the much older K–Ar dates: 4.8 and 4.2 Ma for samples Yug-2 and Yug-3, respectively. It should be noted that the analyzed biotite fractions have anomalously low K content (4.8–5.4%) in spite of a high mineral purity (97–99%). The low K concentration (less than 6%) was also established for biotite fractions of the Kazbek Quaternary lavas [22]. As a rule, the K–Ar dating of these fractions yields ages older than those of the Kazbek lavas, and we doubted the biotite dates for the Amiranisgora Volcano. An attempt to date plagioclase from pink dacites failed, because phenocrysts of this mineral so saturated by atmospheric argon that the radiogenic increment cannot be diagnosed. The groundmass of pink dacites also revealed a considerable contamination by atmospheric argon (94% of the total content; Table 3). The abnormally high contamination of the Amiranisgora pink dacites by atmospheric argon may be a consequence of their specific formation, which caused the iron oxida-

Table 4. Rb-Sr isotopic ages of the samples from the Amiranisgora (Tavshan-Tapa) Volcano

Sample no.	Rock	Fraction	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr \pm 2\sigma$	$^{87}Sr/^{86}Sr \pm 2\sigma$
Yug-2	Pink dacite	Plagioclase	2.8	1620	0.00500 ± 0.00005	0.704094 ± 0.000015
		Amphibole	7	206	0.0960 ± 0.0007	0.704016 ± 0.000014
		Whole rock	70	585	0.3480 ± 0.0008	0.704062 ± 0.000015
		Groundmass	83	443	0.5418 ± 0.0013	0.704044 ± 0.000023
		Biotite	277	68	11.81 ± 0.03	0.704895 ± 0.000016
Yug-3	Gray dacite	Plagioclase	2.1	1667	0.00360 ± 0.00006	0.704122 ± 0.000015
		Amphibole	10	100	0.2859 ± 0.0008	0.704115 ± 0.000013
		Matrix	88	660	0.3888 ± 0.0010	0.704143 ± 0.000011
		Biotite	210	103	5.866 ± 0.023	0.705075 ± 0.000014

tion and appearance of abundant rock defects and micropores.

Almost all mineral components of the Amiranisgora dacites were dated by the Rb–Sr method (Table 4). Owing to presence of biotite, the mineral with high Rb/Sr ratio, we expected that the Rb–Sr dates may clarify significance of various groups of the K–Ar dates.

The Rb–Sr isotope analyses of different components and whole rock samples of the dacites (Yug-2 and Yug-3) showed that their Sr isotope composition is misbalanced. A similar phenomenon was reported for the Quaternary lavas of the Kazbek area and the Elbrus volcanic center of Greater Caucasus [28]. The initial $^{87}Sr/^{86}Sr$ ratio in plagioclases and whole-rock samples is more radiogenic than in the groundmass and amphiboles (Fig. 8). As for the biotite, the high content of radiogenic ^{87}Sr accumulated *in situ* "shades" to some extent the initial $^{87}Sr/^{86}Sr$ ratio in this mineral.

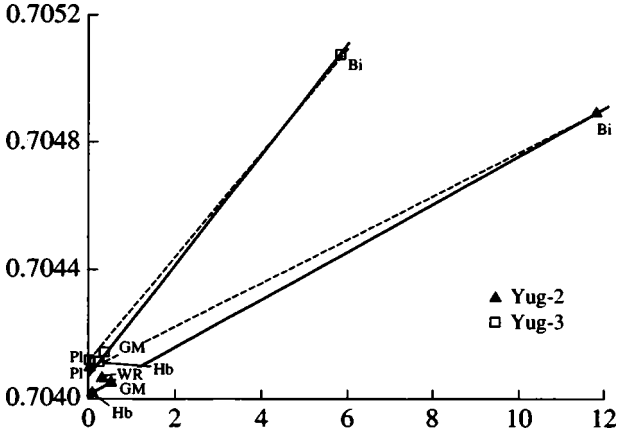

Fig. 8. Rb–Sr isochron diagrams for samples Yug-2 and Yug-3 (the Amiranisgora dacites). Sample Yug-2: Bi–GM–Hb isochron, age 5.3 ± 0.1 Ma, $I_0(Sr)$ = 0.704008 ± 0.000006, MSWD = 0.24; Bi–Pl isochron, age 4.78 ± 0.13 Ma, $I_0(Sr)$ = 0.704094 ± 0.000015. Sample Yu-3: Bi–GM–Hb isochron, age 12.0 ± 0.2 Ma, $I_0(Sr)$ = 0.704072 ± 0.000009, MSWD = 1.5; Bi–Pl isochron, age 11.4 ± 0.25 Ma, $I_0(Sr)$ = 0.704121 ± 0.000015 (Bi is biotite, GM groundmass, Hb amphibole, WR whole rock, Pl plagioclase).

Provided absence of the initial ^{87}Sr excess in biotite, the three-point isochron (biotite–groundmass–amphibole) for Sample Yug-2 corresponds to age of 5.3 ± 0.1 Ma. The $(^{87}Sr/^{86}Sr)_0$ value of 0.704008 ± 0.000006 calculated for biotite is lesser, in terms of minor measurement errors, than that established for the plagioclase. On the other hand, if assume equal $^{87}Sr/^{86}Sr$ initial ratios for biotite and plagioclase, then the biotite–plagioclase Rb–Sr age for Sample Yug-2 would be 4.78 ± 0.13 Ma. Assuming two possible values of the $^{87}Sr/^{86}Sr$ initial ratio in biotite, we can calculate the possible age range of 4.8–5.3 Ma for Sample Yug-2. The K–Ar biotite age of Yug-2 is close to this range and seems trustworthy. However, the same approach and calculations for Sample Yu-3 yielded the age range of 11.4–12.0 Ma that is completely unreal. Therefore, the Rb–Sr isochron ages, which actually characterize biotite, hardly correspond to the formation time of Amiranisgora dacites. The high Sr concentrations and, consequently, low $^{87}Sr/^{86}Sr$ ratios in biotite from both samples are to be noted. It was mentioned above that the Sr concentrations in all the dacites of the Dzhavakheti Upland greatly exceeded the clarke values.

The isotopic systems of mineral phenocrysts and groundmass in the dacites studied are unbalanced, as it is evident from the Sm–Nd parameters of plagioclase, amphibole, and groundmass (Sample Yug-2 Table 5). In the Amiranisgora dacites, amphibole is enriched in rare earth elements, whereas plagioclase is depleted. The Sm/Nd ratio expressed as $^{147}Sm/^{144}Nd$ isotopic ratio (0.06–0.16) varies significantly, and this is a consequence of rare earth elements fractionation during the dacitic magma crystallization.

Table 5 shows that amphibole and groundmass of dacite Yug-2 show similar Sm–Nd isotopic characteristics: $^{143}Nd/^{144}Nd$ = 0.51277–0.51281 (ϵ_{Nd} = 2.6–3.4), whereas the Nd isotopic parameters of plagioclase are shifted, like Sr concentrations, closer to the crustal values.

Table 5. Sm-Nd isotopic characteristics of phenocrysts and groundmass, Sample Yug-2 (the Amiranisgora pink dacite)

Sample	Nd, ppm	Sm, ppm	$^{147}Sm/^{144}Nd \pm 2\sigma$	$^{143}Nd/^{144}Nd \pm 2\sigma$	ϵ_{Nd}
Plagioclase	2.1	0.22	0.06480 ± 0.00006	0.51269 ± 0.00009	1.0
Groundmass	16	2.3	0.08590 ± 0.00002	0.512771 ± 0.000020	2.6
Amphibole	28	7.6	0.16310 ± 0.00006	0.512814 ± 0.000009	3.4

Thus, the geochronological data obtained show a complicated behavior of three isotopic systems (K–Ar, Rb–Sr, and Sm–Nd) in acidic volcanics. Therefore, the age of the Amiranisgora extrusive can be roughly estimated in the range of 3–5 Ma. There are two significantly different (with analytical errors taken into account) groups of dates falling into the intervals of 2.7–3.2 Ma (K–Ar ages for groundmass) and 4.2–5.3 Ma (K–Ar ages for biotite fractions and Rb–Sr ages for Sample Yug-2). The more precise dating of the Amiranisgora Volcano requires additional investigations. In our opinion, the K–Ar ages of groundmass are more reliable than those of biotite, because of the anomalous chemical composition of the latter (low K and Rb contents against high Sr concentration), the possible presence of excessive argon in biotite phenocrysts, as established previously, and unreliably old Rb–Sr age of biotite from Sample Yug-3. Most likely, the Amiranisgora dacites are 2.7–3.2 Ma old (late Pliocene). At that time, a weak impulse of magmatic activity in the Dzhavakheti Upland resulted in origin of some extrusive structures, e.g., of the Chikiani and Amiranisgora volcanoes composed of acidic and moderately acidic rocks of variable chemical composition. However, the older age of 4.2–5.0 Ma should not be excluded for the Amiranisgora extrusion, because the K–Ar isotope system in dacite groundmass might be disturbed, when it was subsequently heated by the Pliocene basaltic flows, resting on the volcano slopes. In this case, the dacite crystallization time is indicated by the K–Ar age of biotite, the isotopic system of which is more stable.

The Inyak-Dag Volcano

The basaltic andesite flow of the Inyak-Dag Volcano occupies a single hypsometric level within the Akhalkalaki Formation basalts. The K–Ar age (2.05 ± 0.10 Ma) of the basaltic andesite (Yug-123, Table 3) is close to that of the Akhalkalaki basalts (2.9–1.8 Ma, [14]). According to the chemical composition, geological position and age, the Inyak-Dag Volcano and its lava flows can be reliably referred to the plateau basalts (Akhalkalaki Formation) of the Dzhavakheti Upland

The Kir-Dag Volcano (Dzhavakheti Ridge)

The close K–Ar date (2.25 ± 0.1 Ma) was obtained for weakly porphyritic dacite flow of the Kir-Dag Volcano (Sample Yug-118, Table 4). The Kir-Dag Volcano is located on the eastern slope of the Dzhavakheti Ridge. The lava flow forms a low range between the Tsalka and Gomareti plateaus composed of the Akhalkalaki basalts. Its terminal part overlies the Khrami granitoid massif producing alterations at the contact. The dacitic lavas of the Dzhavakheti Ridge were previously considered as older than the plateau basalts ([11] and others). However, the dates obtained indicate that the Kir-Dag dacites are of the same age as the Akhalkalaki Formation basalts. We interpret the Dzhavakheti dacites, including the Kir-Dag rocks, as derivates of basaltic melts of deep-seated origin, or as products of magma mixing with old crustal material, e.g., with rocks of the Paleozoic Khrami massif exposed in the nearest proximity of the Kir-Dag Volcano. The Kir-Dag dacites were likely formed at the terminal stage of the Pliocene magmatic cycle in the Dzhavakheti volcanic area. The stage was characterized by volcanism of basaltic to basaltic andesite composition. The objects of interest for further isotopic-geochronological investigations are dacitic lavas of other volcanoes of the Dzhavakheti Ridge, which are supposed to be coeval with the Kir-Dag lavas.

Volcanoes of the Samsari Ridge

Unexpected results of K–Ar dating were obtained for seven samples (Yug-66, Yug-67, Yug-68, Yug-69, Yug-72, Yug-75, Yug-76) of amphibole–plagioclase dacites from Didi-Abuli, Samsari, and Shavnabada volcanoes of the Samsari Ridge. As noted above, rocks of these volcanoes were considered as members of the Mio-Pliocene Goderdzi Formation buried under basalts of the Akhalkalaki Formation. However, the dates obtained indicate with confidence the Late Quaternary age of dacitic lavas from the Samsari Ridge. The dates for groundmass of the studied samples vary in a narrow range from 180 to 320 ka (Table 3). Probably, the K–Ar ages would be even in a narrower range, if the rocks were not so strongly contaminated by atmospheric argon (96–99% of the bulk argon). This prevents the radiogenic argon to be more precisely determined and, as a consequence, we obtain the increased range of the K–Ar dates.

Interestingly, the age of pyroclastic rocks of the large Samsari caldera (Yug-72, 0.20 ± 0.04 Ma) is identical to that of the dacitic lavas forming the volcano itself (0.18–0.24 Ma). This is compatible with a gradual transition between the pyroclastic facies and dacitic lavas [11]. In addition, gray and pink varieties of the Samsari dacitic lavas yield similar dates being almost

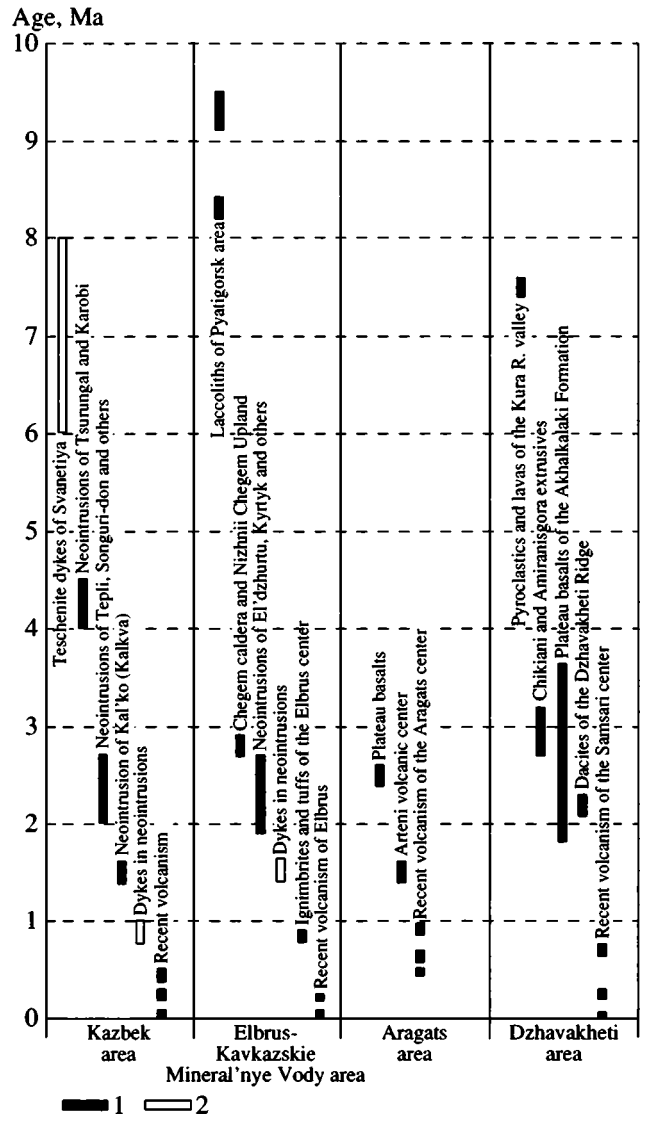

Fig. 9. Geochronology of the Neogene–Quaternary magmatism in some regions of the Caucasian segment of the Alpine belt (this work and data of Arakelyants *et al.* [6]; Borsuk *et al.* [5]; Hess *et al.* [7]; Kostitsyn and Kremenetskii [34]; Ferring *et al.* [14]; and Chernyshev *et al.* [1]): (1) proved and (2) supposed (single dates) stages and phases of magmatic activity (column lengths correspond to stage durations).

equally saturated by atmospheric argon. In distinction, the pink lavas of the Kazbek area appeared unsuitable for the K–Ar dating, as they are contaminated by atmospheric argon to the much greater extent. The Quaternary age of the Samsari volcanoes is also evident from paleomagnetic records studied by Stankevich [19], which are similar to those in recent lavas of the Kazbek area and Elbrus center [29]. The Samsari volcanic structures are weakly eroded, and there are even crater remnants at some mountaintops (e.g., at the Didi-Abuli top). As mentioned above, the assignment of the Samsari volcanics to the Mio-Pliocene Goderdzi Formation [11] was previously doubted by Milanovskii and Koronovskii [4] and by other authors. Nevertheless, the idea of Skhirtladze [11] was commonly accepted after discussion.

In distinction from the Dzhavakheti Ridge predominantly composed of Pliocene basaltic lavas, among which local acidic eruptions took place about 3–2 Ma ago (the Chikiani, Kir-Dag and other volcanoes), volcanoes of the Samsari Ridge originated, according to K–Ar dates, at the final, Quaternary stage of magmatism in the Lesser Caucasus. Thus, it is evident that submeridional deep-seated faults, to which the Dzhavakheti and Abul-Samsari ridges are confined, became activated in different times.

Hyalodacites of the Kechut Formation

The K–Ar dating of hyalodacite (Sample Yug-63) from lava flow of the Godorebi Volcano that is referred to the so-called Kechut Formation implies the late Neopleistocene and even Holocene age (30 ± 30 ka) of this structure. As noted above, the extremely young age of hyalodacites is evidenced by many geomorphological features. These lavas usually erupted from parasitic fissures like those on slopes of the Samsari Ridge volcanoes and less frequently from the main crater, e.g., in the Tavkvetili Volcano in the northern part of the Samsari Ridge. It seems rational to apply other methods (e.g., of radiocarbon dating) in addition to K–Ar age determinations in order to get a deeper insight into the latest stage of volcanism in the Dzhavakheti area. The radiocarbon dating is applicable for fossil remains, which document breaks between lava eruptions and may be found.

The Babakhngo Volcano ("Western Range" of the Samsari Ridge)

The K–Ar dating of dacite (Yug-73, Table 4) from the Babakhngo Mountain yielded an interesting result. This volcanic structure is an element of a low mountain range stretching along the western slopes of the Samsari Ridge and locally separated from it by intermontane depressions, where villages of Samsari, Olaverdi, Buzabeti, Abuli and others are located. The range restricts distribution of basalts of the Akhalkalaki Formation within the synonymous plateau (Fig. 1) and probably is associated with the submeridional fault zone. The K–Ar date of 0.76 ± 0.08 Ma for the dacite groundmass corresponds to the Eopleistocene–Neopleistocene boundary time. The date is the only one and needs to be verified after additional analysis of rocks from the "western range." It is noteworthy that this phase of volcanism (0.76 ± 0.08 Ma) in the Dzhavakheti area, which was caused probably by activation of submeridional fault under the range, is correlative with the activity impuls in the adjacent Aragats area about 700 ka ago [16]. According to our data, the Aragats Volcano originated at the same time.

Phases of the Quaternary Volcanism of the Dzhavakheti Area

Thus, the K–Ar dating of dacitic lavas of the Samsari Ridge indicates their Late Quaternary age. The revealed three phases of volcanism took place in the early Neopleistocene (800–700 ka), middle Neopleistocene (300–200 ka), and late Neopleistocene–Holocene (younger than 50 ka). The last two phases coincide with periods of volcanic activity in the Kazbek area and Elbrus center of Greater Caucasus [8]. The coincidence suggests synchronism of recent volcanic manifestations and interrelation between volcanic impulses in the Greater and Lesser Caucasus. However, in the Aragats area bordering the Dzhavakheti area in the south, the volcanic activity terminated about 450 ka ago [16]. On the contrary, some active volcanoes of the eastern Anatolia (e.g., the Nemrut cone) erupted in the historic time. Consequently, the Quaternary volcanism in the Lesser Caucasus led central eruptions, and the irregular volcanic impulses were of variable intensity in different volcanic areas.

According to the results obtained, the Samsari Ridge is a young volcanic center, and it was incorrect to attribute dacitic lavas and pyroclastic rocks of the ridge to the upper Miocene Goderdzi Formation. Nevertheless, the Goderdzi Formation rocks may be locally exposed within the Samsari Ridge, since even Paleogene deposits crop out here in some erosion windows. Additional investigations of the Dzhavakheti area are needed to detect here rocks of that formation and to determine the easternmost distribution limit of the Goderdzi pyroclastic rocks.

The problem of independent Kechut Formation composed of hyalodacite lavas appears to be a separate object of discussion on the Samsari Ridge geology. It seems reasonable to distinguish a single Late Quaternary volcanic complex of the Samsari Ridge, which is composed of two-pyroxene–amphibole–plagioclase dacitic lavas (Phase I of volcanic activity), amphibole–plagioclase dacitic lavas (Phase II), and the hyalodacite lavas (Phase III).

The dates obtained and geomorphological manifestations of the recent, possibly Holocene volcanism imply that the central part of the Dzhavakheti Upland is an area of potential seismic hazard. The further works should be oriented toward the comprehensive isotopic-geochronological study of the Late Quaternary volcanism in the Samsari Ridge in order to understand the periodicity of eruptions and the variations in volume and composition of erupted material during the volcanism evolution. This study will be of a great prognostic importance for the entire Caucasian region.

The K–Ar Isotopic Systematics for Plagioclase Phenocrysts of the Samsari Ridge Quaternary Lavas

Studying the Quaternary dacitic lavas of the Samsari Ridge, we continued our work on the K–Ar isotope systematics for phenocrysts in young volcanic rocks, which are usable for isotopic dating. The K–Ar isotopic ages were determined for groundmass and plagioclase phenocrysts of three samples (Yug-67, Yug-76, and Yug-73; Table 4). The plagioclase dates are of a wider range (from 1.2 to 8.2 Ma) and much older than the groundmass dates, which more reliably define the formation time of dacitic lavas. Similar results were formerly published [8] for volcanic rocks of the Kazbek area and Elbrus center, which yielded the essentially older K–Ar dates plagioclase phenocrysts in the late Neopleistocene lavas (1.5 Ma) and Eopleistocene ignimbrites (15 Ma). Along with the radiogenic argon excess, plagioclase phenocrysts of all studied volcanics from the Greater Caucasus contain abnormally high concentration of radiogenic ^{87}Sr, as compared with the other rock fractions [28]. The same is typical of the Pliocene dacites of the Amiranisrora Volcano. In discussion on possible causes of this phenomenon, there were mentioned the occurrence of plagioclase xenocrysts [30], the excessive argon capture during phenocryst crystallization [23], and some other factors [28]. In our opinion, the argon excess in phenocrysts may be produced by the following mechanism. Being heated by melts, rocks surrounding intermediate magmatic chambers lose a considerable amount of radiogenic argon. The argon partial pressure increased in the magmatic fluid phase may result in both the argon saturation in the liquid phase and the argon capturing by phenocrysts during their crystallization. After eruption, the degassing magma loses the argon excess, whereas mineral phenocrysts retain a part of it. If this hypothesis is correct, the argon excess in phenocrysts should be dependent to the great extent on composition and age of the host rocks. In addition, different mineral phases are likely able to capture variable amounts of excessive argon at high temperatures. For example, analysis of the Elbrus volcanics showed that the ^{40}Ar excess in clinopyroxene is only 0.02 ppb against 0.86 ppb in plagioclase with a higher K content [31].

Thus, summarizing the available data on the K–Ar isotope systematics for plagioclase phenocrysts in the young Caucasian volcanics, we should conclude that this mineral is unsuitable for isotope dating of most Neogene–Quaternary volcanic rocks.

CONCLUSION

The isotopic-geochronological investigations of moderately acidic Neogene–Quaternary magmatic rocks from the Dzhavakheti area of the Lesser Caucasus showed that one more area of young volcanism in the Caucasian segment of Eurasia, namely the Samsari volcanic center, is of the Late Quaternary, but not of the early Pliocene age, as it was formerly assumed. The K–Ar dates outline three discrete phases of volcanic activity in the study region: the formation of "western range" volcanoes (800–700 ka), the origin of Samsari caldera and other volcanoes in the main ridge (300–

200 ka), and the formation of parasitic volcanoes and hyalodacite lava flows (earlier than 50 ka). The dates obtained evidence for the volcanic activity in the Samsari Ridge until the late Neopleistocene and possibly Holocene. Accordingly, this part of the Dzhavakheti area is potentially hazardous one.

The isotopic-geochronological and petrochemical characteristics of the studied dacitic lavas from the Dzhavakheti Upland imply that the Mio-Pliocene Goderdzi Formation in its former scope is invalid. The late Miocene volcanics (7–8 Ma old) of the Dzhavakheti Upland are represented only by pyroclastic units with intraformational dacitic flows, which constitute the lower parts of exposures in the Kura and Paravani river valleys. The extrusive domes of the Chikiani and Amiranisgora volcanoes composed of subalkaline obsidians and biotite–amphibole dacites were formed in the Pliocene time. Dacitic lavas of the Dzhavakheti Ridge erupted in the late Pliocene, simultaneously with plateau basalts widespread in the Transcaucasia. The Samsari volcanic center appeared at the final, Neopleistocene stage of volcanism and was active until the late Neopleistocene–Holocene time.

Thus, the total duration of Neogene–Quaternary magmatic activity in the Dzhavakheti Upland was about 7 to 8 m.y. All the dacitic volcanics, regardless their ages, belong to the calc-alkaline series with moderate potassium content. Some stages and phases of volcanic activity in the Dzhavakheti Upland are well correlative with concurrent events in adjacent volcanic areas of the Greater and Lesser Caucasus [31, 32] (Fig. 9).

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Biostratigraphy and Structure of the Upper Quaternary Deposits and Some Paleogeographic Features of the North Caspian Region

Yu. P. Bezrodnykh*, B. F. Romanyuk*, S. V. Deliya, R. D. Magomedov**, V. M. Sorokin***, O. B. Parunin***, and E. V. Babak******

** Joint Stock Company Morinzhgeologiya, Riga, Latvia*
*** Open Joint Stock Company Lukoil-Astrakhan'morneft' Ltd, Astrakhan', Russia*
**** Moscow State University, Moscow, Russia*
e-mail: sorokin.msu.ru
***** Presidium of the Russian Academy of Sciences, Moscow, Russia*
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Abstract—Among materials of the OJSC Lukoil obtained in the northern Caspian region, there are cores samples recovered from the depth of 7 to 37 m below bottom, piston cores, and high-resolution seismic profiles. These materials are used to measure ^{14}C age of sediments and to study biostratigraphy and structure of the Upper Quaternary deposits and paleogeography of the region. Mollusks, benthic foraminifers, diatoms, spores and pollen from sediments recovered by a series of boreholes characterize in detail and substantiate the discrimination and correlation of the Novocaspian, Mangyshlak, upper and lower Khvalynian, Atel', and upper Khazarian beds. The ^{14}C dates indicate that the beds accumulated in the time intervals of 0–7000, 7000–9000, 9000–16000, 17000–30000, and >30000 years, respectively. Lithologic and seismoacoustic data suggest four large sea-level drops relative to the current level in the time spans of the Mangyshlak (–40 to –50 m), Enotaevsk (–20 m), Atel' (–70 m), and the Chernyi Yar–Astrakhan' (or Intra-Khazarian) regressions (–90 m). The events are recognizable based on deep incision channels of the Paleovolga and on lacustrine deltaic and continental facies.

Key words: Caspian Sea, biostratigraphy, ^{14}C dating, paleogeography, Late Quaternary, marine sediments.

INTRODUCTION

The Quaternary stratigraphy of the Caspian Sea was investigated based on the sediments exposed or drilled on the adjacent land. The problem of stratigraphic units discrimination, their faunal characteristics, and ^{14}C dating, structure and paleogeography of the Caspian sediments, as well as the overview of available hypotheses are completely presented in papers by Fedorov [1] and Svitoch and Yanina [2]. Several stages, the Bakuan, Khazarian, Khvalynian, and Novocaspian, were distinguished in the Quaternary history of the basin. The transgression periods of a high sea level are well recorded on land, while the regressive marine deposits are almost unstudied as they are submerged. Accordingly, the Caspian Sea history is fragmentary understood because of scarce biostratigraphic data characterizing the regressive phases.

The uninterrupted geological history of the Caspian Sea can be reconstructed based on continuous sections of marine sediments. However, the most of data available up to now are those referred to shallow-water Novocaspian deposits recovered by piston cores that bear paleontological remains, mainly bivalve shells used to ground the Caspian stratigraphic scale. In addition, there are known scarce records in the Khvalynian and older sediments that occur at a considerable depth beneath the sea floor. Paleontological evidences of the youngest Mangyshlak and ancient regressive phases of the Caspian Sea are extremely scarce, and present concepts are mainly based on the echo sounding data that reveal the terrace-like ledges at the sea bottom [3] and, rarely, on the direct observations of bottom deposits [4]. In general, the observations are ambiguous and often discrepant. All the above is typical of the northern part of the Caspian Sea that is poorly understood in stratigraphic and paleogeographic aspects because of its shallowness.

In the course of the geologic prospecting in the northern Caspian Sea, which is sponsored by OJSC Lukoil, the geological engineering explorations included drilling of boreholes and seismic profiling aimed to provide the safe positioning of the self-hoisting drilling rig. Nineteen boreholes 25 to 100 m deep were drilled and over 300 piston cores up to 3.5–4.0 m long were sampled from the depth of 7 to 36.5 m in 1997–2001 in five areas located in the northern Caspian Sea within its middle zone extending for about 110 km from north to south. The two-frequency seismic profiling was carried out with resolutions of 0.3–0.5 and 1.5–2.0 m for subbottom intervals of 10–15 and 80–100 m,

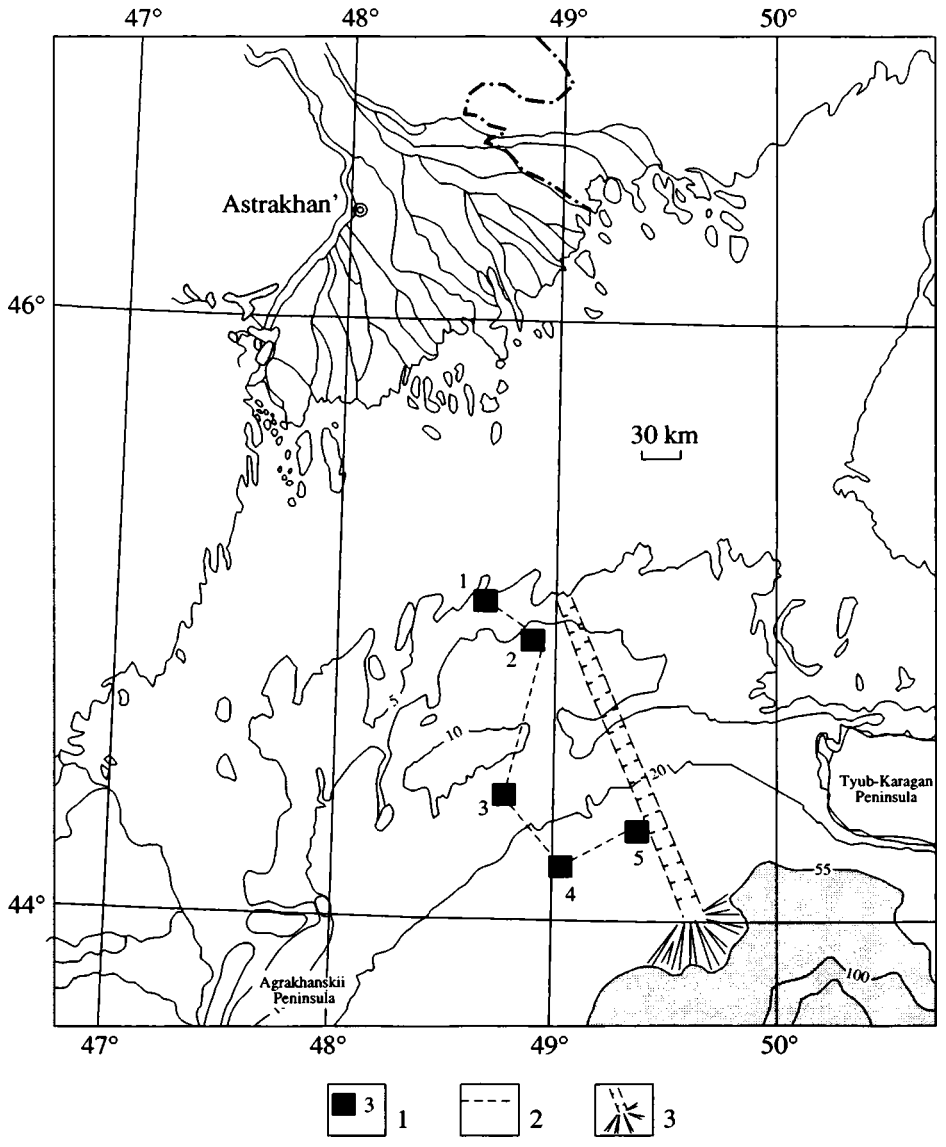

Fig. 1. Locations of studied areas and seismic profiles: (1) studied areas and their numbers; (2) seismic profiles; (3) buried river valley and terminal fan.

respectively. The regional seismic profiles of that time connect areas of regular profiling grids and enable correlation between all the boreholes drilled. The disposition of studied objects and regional profiles is shown in Fig. 1.

The materials obtained characterize the structure and composition of the Upper Quaternary sequence in a vast area of the northern Caspian region. Many continuous sections of the Upper Quaternary sediments are now described. Their facies interrelations and stratigraphy well substantiated geochronological dates, as well as the considered paleogeographic problems, represent an essential contribution to the knowledge of the basin's geology. The recovered sediments were tested for bivalve assemblages and benthic foraminifers (T.M. Maier), diatoms (L.G. Pirumova), pollen and spores (N.O. Rybakova), and biologic composition of organogenic deposits (O.N. Uspenskaya). They were subjected to radiocarbon dating as well.

INVESTIGATION RESULTS

Stratigraphy

Stratigraphic subdivision of the recovered Upper Quaternary sediments is based on the species composition of mollusks (Table 1), diatoms, benthic foraminifers, and other organic remains. The age inferences are confirmed by the ^{14}C dates (Table 2). The discriminated Novocaspian, Mangyshlak, Khvalynian, and Khazarian stratigraphic horizons are characterized below.

Table 1. Mollusk assemblages in the Pleistocene–Holocene sediments of the northern Caspian Sea

Area 1	Area 2	Area 3	Area 4	Area 5
Novocaspian Horizon (upper) (0–1 ka)				
Abra ovata (Philippi), *Mytilaster lineatus* Gmel., *Cerastoderma lamarcki* (Reeve), *Didacna barbotdemarnyi* (Grimm), *D. trigonoides* Pall., *D. parallella* (Bog.), *Hypanis caspia* (Eichw.), *H. leviuscula* (Eichw.), *H. vitrea* (Eichw.), *Dreissena polymorpha caspia* (Pall.), *Dr. polymo-rpha polymorpha* (Pall.), *Unio* sp., *Pyrgula (Clessiniola) variabilis* (Eichw.), *P. (Micromelania) caspia* (Eichw.), *P. (Hydrobia)* sp., *Theodoxus pallasi* Lindch., *Viviparus viviparus* (L.)	*A. ovata, M. lineatus, C. lamarcki, D. barbotdemarnyi, D. trigonoides, H. caspia, H. vitrea, H. leviuscula, Dr. polymorpha caspia, Dr. rostriformis distincta, Unio* sp., *P. caspia, Th. pallasi, V. viviparus.*	*A. ovata, M. lineatus, C. lamarcki, D. baeri* (Grimm), *D. barbotdemarnyi, D. parallella, H. caspia, H. vitrea, H. leviuscula, Dr. polymorpha caspia, Pyrgula, Th. pallasi.*	*D. barbotdemarnyi, D. parallella, D. praetrigonoides Nal, H. caspia, H. vitrea, H. leviuscula, Dr. polymorpha caspia, Dr. rostriformis distincta*	
Novocaspian Horizon (middle) (1–4 ka)				
The same assemblage but missing *Abra ovata* and *Mytilaster lineatus*				
Novocaspian Horizon (lower) (4–7 ka)				
D. barbotdemarnyi, H. caspia, H. leviuscula, H. vitrea, Dr. polymorpha caspia, Dr. polymorpha polymorpha, Unio sp., *P. variabilis, Th. pallasi*	*D. barbotdemarnyi, H. caspia angusticostata* (Borc.), *H. vitrea, H.leviuscula, Dr. polymorpha caspia, Dr. polymorpha polymorpha, Unio* sp., *P. variabilis*	*D. parallella, D.* cf. *cristata, H. caspia, H. vitrea, H. leviuscula, Dr. polymorpha caspia, Dr. polymorpha polymorpha, Dr. rostriformis distincta, Pyrgula, Th. pallasi*	*D. barbotdemarnyi, D. parallella, D. praetrigonoides, H. caspia, H. vitrea, H. leviuscula, Dreissena polymorpha caspia, Dr. rostriformis distincta*	
Mangyshlak Horizon (7–9 ka)				
Dr. polymorpha polymorpha, Hypanis caspia, H.vitrea, H. leviuscula, Unio pictorum (L.,), *Unio* sp., Limneidae				
Khvalynian Horizon (upper) (9–16 ka)				
D. parallella, D. cristata (Bog.), *H. caspia, Dr. polymorpha caspia, Dr. rostriformis distincta, P. caspia, P. elegantula* (Cless. et Dyb.), *P. variabilis, Lithoglyphus (Pseudamnicola) caspius Krin*	*D. parallela, D. cristata, H. caspia, H. plicatus, H. vitrea, Dr. polymorpha polymorpha, Th. pallasi*	*D. praetrigonoides, D. parallella, D. subcatillus Andrus., H. caspia, H. vitrea, Dreissena polymorpha caspia*	*D. barbotdemarnyi, D. praetrigonoides, D. cristata, H. caspia, H. vitrea, Dr. polymorpha caspia, Pyrgula*	*D. barbotdemarnyi, D. praetrigonoides, D. cristata, D. subcatillus, H. caspia, H. vitrea, H. leviuscula, Dr. polymorpha polymorpha, Dr. rostriformis distincta, L. caspius, P. caspia, Th. pallasi*
Khvalynian Horizon (lower) (17–>30 ka)				
D. parallella, D. subcatillus, D. praetrigonoides, H. caspia, H. leviuscula, Dr. polymorpha polymorpha, Dr. polymorpha caspia, Dr. rostriformis distincta, P. caspia, P. variabilis, P. grimmi, Th. pallasi	*D. parallella, D. subcatillus, D. cristata, D. praetrigonoides, D. zhukovi Fed., D. protracta Eichw., H. caspia, H. vitrea, Dr. polymorpha polymorpha, Dr. polymorpha caspia, Dr. rostriformis distincta, P. caspia, P. grimmi, Th. pallasi*	*D. subcatillus, D. cristata, H. caspia, Dr. polymorpha caspia, Dr. polymorpha polymorpha, Dr. rostriformis distincta, P. caspia, P. spica, L. caspius, Th. pallasi*	*D. subcatillus, D. parallella, D. praetrigonoides, D. cristata, D. cf. protracta, H. caspia, H. vitrea, H. leviuscula, Dr. polymorpha caspia, Dr. rostriformis distincta, Pyrgula, Th. pallasi*	*D. subcatillus, D. parallella, D. zhuckovi, D. praetrigonoides, H. caspia, r. rostriformis distincta, P. caspia, P. variabilis, L. caspius, Th. pallasi*

Table 1. (Contd.)

Area 1	Area 2	Area 3	Area 4	Area 5
	Khazarian Horizon (upper)			
Corbicula fluminalis (Muller), *H. caspia.*	*D. baeri* (Grimm) (forms close to *D. surochanica Andrus., D. nalivkini Wass, D. subcatillus, D. parallela, D. protracta*), *Dr. rostriformis distincta, Dr. polymorpha caspia, Dr. polymorpha polymorpha, H. caspia, P. caspia, P. variabilis, P. grimmi, L. caspius, Th. pallasi*			
	Khazarian Horizon (lower)			
	Didacna cf. subcatillus, D. cf. subovlis Prav., H. caspia, Dr. rostriformis distincta, P. caspia, P. grimmi, Th. pallasi			

The Novocaspian Horizon is subdivided based on the mollusk evidence from areas 1–3 and 5 (sea depth from 10–17 to and 36 m) into three beds (Table 1).

The upper bed yields the *Dreissena polymorpha caspia* (Pall.)–*Didacna baeri* (Grimm) assemblage and is clearly distinguishable because of presence of *Mytilaster lineatus* Gmel and *Abra ovata* (Philippi). The bed also bears *Cerastoderma lamarcki* (Reeve), *Hypanis (Monodacna) caspia* (Eichw.), *H. (Adacna) vitrea* (Eichw.), *H. (Adacna) leviuscula* (Eichw.), scarcer *Didacna barbotdemarnyi* (Grimm) and other didacnas, and gastropods *Pyrgula* and *Theodoxus*. The bed's base is about 1 ka old according to radiocarbon dates (Table 2).

The middle bed contains the *Dreissena polymorpha caspia* (Pall.)–*Didacna barbotdemarnyi* (Grimm) assemblage with the marker species *Cerastoderma lamarcki* (Reeve). The assemblage includes *Hypanis (Monodacna) caspia* (Eichw.), slightly scarcer *H. (Adacna) vitrea* (Eichw.), *H. (Adacna) leviuscula* (Eichw.), and rare *Dreissena rostriformis* (Andrus.) in association with didacnas and gastropods. The base of the bed is dated at 3–4 ka.

The lower bed is marked by absence of *Cerastoderma lamarcki* (Reeve). It was accumulated between 3–4 and 6.8 ka ago. In the Area 4, the Novocaspian mollusk assemblage from the depth of about 27 m is of a similar composition, also lacking *Cerastoderma lamarcki*, and the bed's base is about 6 ka old.

Diatoms from the Novocaspian deposits were studied in sediments of Area 2. A single sample bearing diatom remains is referred here to the lower bed. It yielded 32 taxa of freshwater diatoms only. These are the oligohaline species surviving salinity not greater than 5‰, and forms indifferent even to the value of 5‰ dominate among these species and associate with several halophilic taxa tolerant to such a salinity. All the identified taxa of the assemblage are referred to Centrophyceae, genera *Aulacosira* and *Stephanodiscus* included, and Pennatophyceae representing the other 11 genera. The assemblage consists of *Aulacosira italica, A. italica* subsp. *subarctica, A. granulata, Stephanodiscus astraea, Fragilaria bicapitata, F. brevistriata, F. brevistriata* v. *sulcapitata, F. construens, F. construens* v. *subsalina, F. inflata, F. intermedia, F. lanceolata* v. *minuta, F. leptostauron, F. pinnata, F. virescens, F. virescens* v. *capitata, Synedra uena, Cocconeis placentula, C. placentula* v. *euglypta, Eunotia pectinalis* v. *euglipta, Navicula placentula, N. placentula* f. *minuta, N. placentula* f. *rostrata, Pinnularia microstauron, Caloneis bacillum, C. silicula, Girosigma acuminatum, Amphora ovalis, Gomphonema abbreviatum, Nitzschia acuta, N. kuetzingiana*, and *N.* sp.

Centrophyceae forms are characteristic of the river and lacustrine plankton, and their representatives *Aulacosira italica* subsp. *subarctica* and *A. granulata* prefer cool waters. The genus *Fragilaria*, inhabitant of the intertidal zone in stagnant basins, stands out among the other benthic diatoms because of species diversity and abundance. Most species of this genus are epibiontic and often dwell on watered stacks and in mosses. Other diatoms of the assemblage are also widespread in intertidal zones of different basins as components of benthos or epibionts. Species *Eunotia pectinalis* var. *minor* inhabits bogs, damp mosses, and watered stacks. *Navicula* forms are inhabitants of the intertidal zone of alkaline basins. *Pinnularia microstauron* prefers alkaline or weakly acidic flowing and stagnant waters. *Caloneis* species are widespread on the bottom of different basins. *Girosigma acuminatum* dwells on the bottom and rarely in planktonic communities of various basins.

The investigation of spores and pollen showed their extremely irregular distribution in the section, from complete absence to relative abundance. The lower Novocaspian bed sampled in Area 2 is found to be the most fossiliferous. It yields 130 well-preserved palynomorphs. The pollen spectrum includes 25.4% of arboreal taxa, 64% of herbaceous and 10.6% of spores. The

Table 2. Radiocarbon dates of the northern Caspian Sea sediments

Sample	Borehole (station), interval (m)	Material	Age (years)
	Area 1		
MSU-1566	Station 41, 42	Shells	Q_{IV} nk (3200 ± 50)
MSU-1567	Station 53 (0.4–0.5)	Shells	recent
MSU-1570	Station 36 (1.2–1.7)	Shells	Q_{IV} nk (4130 ± 70)
MSU-1571	Station 36 (0.8–1.05)	Shells	Q_{IV} nk (2895 ± 60)
MSU-1572	Station 36 (0.2–0.48)	Shells	Q_{IV} nk (860 ± 40)
MSU-1587	IGS-3 (24.5–25.0)	Shells	Q_{III} hv_1 (18330 ± 150)
MSU-1588	IGS-3 (21.1–21.2)	Shells	Q_{III} hv_1 (21460 ± 200)
MSU-1589	IGS-3 (9.0–10.1)	Shells	Q_{III} hv_1 (17645 ± 160)
MSU-1590	IGS-1 (6.5–7.5)	Shells	Q_{IV} nk (680 ± 30)
	Area 2		
MSU-1493	Station 70, 69, 60, 18, 10, 9 (1.15–1.65)	Organic matter	Q_{IV} mg (9420 ± 60)
MSU-1494	Station 49 (3.38–3.55)	Organic matter	Q_{IV} mg (9860 ± 240)
MSU-1495	Station 19 (2.3–2.45)	Organic matter	Q_{IV} mg (9860 ± 330)
MSU-1496	Station 68 (2.4–2.5)	Organic matter	Q_{IV} mg (9300 ± 110)
MSU-1556	IGS-3 (21.0–22.5)	Shells	Q_{III} hv_1 (27200 ± 340)
MSU-1559	Station 38 (1.0–1.15)	Shells	Q_{IV} nk (6830 ± 60)
MSU-1561	Station 3 (1.68–1.77)	Organic matter	Q_{IV} mg (7630 ± 440)
MSU-1562	IGS-3 (8.2–9.1)	Shells	Q_{III} hv_2 (6620 ± 130)
	Area 3		
MSU-1507	IGS-3 (21.6–21.8)	Shells	Q_{III} hv_2 (16900 ± 120)
MSU-1508	Well 1 (14.7–14.9)	Shells	Q_{III} hv_2 (9230 ± 165)
MSU-1509	Station 35 (2.95–3.05)	Shells	Q_{IV} nk (6610 ± 60)
MSU-1510	Station 25 (1.0–1.55)	Shells	recent
MSU-1511	IGS-3 (3.1–3.5)	Shells	Q_{IV} nk (1330 ± 60)
MSU-1512	Station 12 (3.25–3.35)	Shells	Q_{IV} nk (4780 ± 50)
MSU-1513	Station 31 (0.66–0.88)	Shells	recent
	Area 4		
MSU-1555	IGS-2 (21.7–21.9)	Shells	Q_{III} hv_1 (21090 ± 320)
MSU-1557	IGS-2 (14.6–14.7)	Shells	Q_{III} hv_2 (16900 ± 110)
MSU-1558	IGS-2 (7.75–7.85)	Shells	Q_{IV} mg (8540 ± 70)
MSU-1560	Station 38 (0.8–1.05)	Shells	Q_{IV} nk (1690 ± 40)
MSU-1563	Station 54 (0.9–2.4)	Shells	Q_{IV} nk (5750 ± 80)
	Area 5		
MSU-1591	IGS-1 (29.4–29.6)	Shells	Q_{III} hv_1 (22190 ± 400)
MSU-1592	IGS-1 (16.9–17.2)	Shells	Q_{III} hv_1 (19325 ± 175)
MSU-1593	IGS-1 (15.1–15.2)	Shells	Q_{III} hv_1 (16075 ± 120)
MSU-1594	IGS-1 (11.0–11.5)	Shells	Q_{III} hv_2 (15710 ± 170)
MSU-1595	IGS-3 (4.8–4.9)	Shells	Q_{III} hv_2 (12870 ± 100)
MSU-1597	IGS-1 (37.6–37.9)	Shells	Q_{III} hvv_2 (30150 ± 610)
MSU-1598	IGS-1 (22.3–23.7)	Shells	Q_{III} hv_1(30085 ± 600)
MSU-1599	IGS-1 (19.3–19.4)	Shells	Q_{III} hv_1 (>30000 ± 600)

pine pollen makes up the bulk of arboreal taxa and represents *Pinus* s/g *Haploxylon* (3.7%), *P.* s/g *Diploxylon* (1.5%), and *Pinus* sp. (3%). The birch pollen is rather common (14.7%). Large well-preserved pollen grains occur in association with smaller, light, and slightly crumpled pollen. Additionally we recorded *Salix* (1.5%) and a strongly deformed pollen grain of *Picea* (0.7%).

The herbaceous and fruticose pollen is markedly diverse. It primarily represents *Artemisia* (20%), Gramineae (19.2%), and Chenopodiaceae (6.2%), which are accompanied by Cyperaceae (1.5%), Liliaceae, Ranunculaceae, Compositae, Umbelliferae, Caryophyllaceae, Rubiaceae, Polygonaceae, etc. (up to 17% of the spectrum in sum). Among spores, *Bryales* constitutes 10% and *Sphagnum* 0.7%. It should be noted that small-sized green mosses commonly grow beneath the grass cover of meadow herbs, and presence of their spores in the spectra dominated by herbaceous pollen is quite explicable.

The Mangyshlak Horizon yields an impoverished *Dreissena polymorpha polymorpha* (Pall)–*Hypanis (Monodacna) caspia* (Eichw.) assemblage of mollusks that shows admixture of freshwater species and lacks *Didacna* forms. This assemblage is characteristic of the freshened waters of the modern Caspian Sea. The composition of benthic foraminifers, especially a great abundance of *Mayerella brotzkajae* (Mayer), also indicates the decreased salinity.

Five samples from Area 2, which are rich in peat- and sapropel-like matter, were tested for plant remains and other organisms. In the group microscopic biotic remains, higher plant taxa constitute 9.0–45.5%, diatoms 43.5–72.4%, yellow-green algae 1–5.6%. Green, blue-green, and clorococcalean algae are as abundant as 5.6–28.1%, and sponges represent 0.3–5.2%.

Macroremains of higher plants macerated by washing through the 0.25-mm sieve are represented by cane *Phragmites communis* (0.7–22.5%), *Typha* (0.5–1.5%), hornwort *Ceratophyllum demersum* (0.5–93.5%), Nymphaeaceae (0.5%), Gramineae, roots, caulis, and epidermis of herbaceous plants (up to 22.5%), and by the "mummified" remains of unrecognizable structure (0–75.0%). The diatoms are mainly represented by benthic epibionts inhabiting freshwater basins. Certain samples contain the insect, mollusk, and ostracode remains.

The composition of studied remains indicates the deltaic and coastal brackish- to freshwater settings overgrown with higher plants. These were areas of purer water with variable content of carbonates and organic matter, and with reducing conditions near the bottom, where deposits were enriched in iron sulfides, e.g., in hydrotroilite, and had admixture of gypsum crystals.

The composition of spores and pollen was identified in two samples of organogenic deposits from Area 2. The samples considerably differ in abundance of pollen and spores. One sample is rich in organic matter that consists of cells and fragments of plant tissue, fossil fungi remains, etc. Among 120 palynomorph taxa, the herbaceous and shrub pollen constitutes 56.7%, spores 42.5%, and arboreal pollen (*Betula*) 0.8%. The herbaceous group includes Gramineae (17.5%), Chenopodiaceae (18.3%), *Artemisia* (12.5%), and herbs (8.4%) with representatives of Liliaceae (1.8%), Umbelliferae (1.8%), Ranunculaceae (0.8%), Saxifragaceae (0.8%), Sparganiaceae (1.6%), and others. Spores belong to green mosses *Bryales* (41.7%) and ferns Polypodiaceae (0.8%). The second sample contains a great amount of spores and pollen. Among 510 palynomorphs, there are xerophyte grass pollen (78.2%) and spores (20%), which make up the bulk of the spectrum. The arboreal pollen is represented by pine (0.8%), birch (0.8%), and oak (0.2%). The herbaceous pollen is abundant and diverse. This group represents *Ephedra* (0.8%), Gramineae (13%), Chenopodiaceae (42%), *Artemisia* (7%), and herbs (16.2%). Among the latter, there are Umbelliferae (0.4%), Liliaceae (2.2%), Cruciferae (0.6%), Ranunculaceae (0.6%), Labiatae (0.4%), Compositae (9.6%), Leguminosae, Caryophyllaceae, Campanulaceae, Rosaceae, Polygonaceae, and Saxifragaceae (1.2%). Spores belong to green mosses (19.4%) and ferns Polypodiaceae (0.6%). Palynological assemblages from both samples characterize open landscapes almost lacking arboreal plants and predominance of steppe xerophytes. Abundant spores of green mosses characteristic of peat are explicable and do not point to humid conditions. The climate was rather arid.

The Khvalynian Horizon is distinguishable in the section because of its peculiar *Didacna* species: *Didacna praetrigonoides* Nal., *D. parallella* (Bog.), *D. cristata* (Bog.), *D. subcatillus* Andrus., *D. protracta* Eichw., and *D. zhuckovi* Fed. In all studied areas it is subdivided into two subhorizons.

The upper Khvalynian Subhorizon yields *Didacna praetrigonoides* Nal., *D. cristata* (Bog.), *D. parallella* (Bog.), *D. subcatillus* Andrus., and *D. barbotdemarnyi* (Grimm) in addition to *Hypanis* and *Dreissena* forms dominating in shallower areas 1 and 2. The radiocarbon dates suggest that accumulation of corresponding sediments took place in the time span from about 9 to 15–16 ka. The lack of benthic foraminifers, ostracodes, and diatoms can indicate either a significant depth, or the unfavorable conditions for their burial.

In the palynological assemblage of the upper Khvalynian sediments from Area 2, herbaceous pollen represents 91%, spores, 6%, and arboreal pollen only 3% of the spectrum. The xerophilous grasses like Chenopodiaceae (51%) and *Artemisia* (28%) absolutely predominate among the herbaceous plants. Herbs represented by Liliaceae, Ranunculaceae, and others, constitute 12% of the spectrum. Among the arboreal plants *Pinus* sp. (1.6%) and *Picea* (1%) were recorded. Spores of

Bryales (6%) and Polypodiaceae (0.6%) occur in association with ancient redeposited forms.

The lower Khvalynian Subhorizon is characterized by abundance of *Didacna* forms, which dominate in places over *Dreissena* and *Hypanis* species. Widespread among them are smaller *Didacna parallella* (Bog.), *D. subcatillus* Andrus., and their transitional forms, while *D. cristata* (Bog.), *D. protracta* Eichw., *D. zhukovi* Fed., and *D. praetrigonoides* Nal. are scarcer. The deposits are older than 17 ka, and some dates are over 30 ka. The subhorizon lacks diatoms, ostracodes, and benthic foraminifers like the upper Khvalynian Subhorizon.

In the palynospectrum from an upper part of the lower Khvalynian sediments (Area 2), the arboreal pollen is represented by a single grain of *Pinus* sp. The spectrum almost entirely consists of non-arboreal pollen of predominant Chenopodiaceae (52%) and *Artemisia* (16.4%). It also includes Gramineae (5.5%) and herbs (12.9%) including Liliaceae, Ranunculaceae, Rosaceae, Saxifragaceae, etc., and Cyperaceae (3%). The *Bryales* spores constitute 9.7% of the assemblage. Sediments also contain the fungi remains and redeposited forms, among which there is Permian *Triatopinites*.

The Khazarian Horizon. The Khazarian deposits bearing faunal remains were recovered in areas 1 and 2. In Area 1, numerous *Corbicula fluminalis* (Muller) and rare *Hypanis (Monodacna) caspia* (Eichw.) were identified. The sediments recovered in Area 2 yield a more diverse faunal assemblage. The most interesting forms of the latter are larger *Didacna nalivkini* Wass and *D. baeri* (Grimm), morphologically very close to *D. surochanica* Andrus., which coexist with smaller *D. parallella* Bog. Also typical of the assemblage are larger *D. subcatillus* Andrus. and more diverse gastropods. In Area 2, the oldest sample bearing fauna (interval 38.5–39.4 m) yielded large incomplete shells of *D.* cf. *subcatillus* Andrus. and *D.* cf. *subovalis* Prav. The above molluscan assemblages indicate at least the late Khazarian age of the recovered sediments. However, they can be assigned to the lower Khazarian as well based on the last species.

In the palynospectrum of the upper Khazarian deposits from Area 2, the arboreal taxa constitute 23%, herbaceous forms 54%, and spores 23%. The arboreal pollen group includes *Pinus* sp. (11%), *Betula* (9%), *Alnus* (1%), and *Corylus* (2%). The bulk of herbaceous pollen belongs to Chenopodiaceae (39%), Gramineae (5%), and *Artemisia* (3%). Herbs, among which there are Ranunculaceae, Compositae, and others, make up 7% of the spectrum. Spores are referred to green moss *Bryales* (17%), *Sphagnum* (4%), and Polypodiaceae (2%). The redeposited Mesozoic *Glechenia* spores are also recorded. Pollen, especially that of arboreal plants, is poorly preserved in the sample. However, the spectrum can be compared with that of Novocaspian deposits based on abundance of arboreal pollen and on presence of sphagnum mosses in the spore group. Assemblages of this kind are likely characteristic of a moderately humid climate.

The spectrum of older beds is quite different, containing arboreal (1.8%) and herbaceous (85.5%) pollen, and spores (12.7%). A single grain of *Betula* was found in the arboreal plant group. The herbaceous plants are represented by *Artemisia* (9%), Chenopodiaceae (54.5%), Gramineae (1.8%), Compositae (1.8%), Ranunculaceae (16.3%), and others. The spore group includes *Bryales* (9%) and *Sphagnum* (3.7%). Pollen is well preserved, and redeposited, mainly Mesozoic, spores were encountered.

Structure and Composition of Late Quaternary Sediments

The generalized lithologic sections with sampled levels and main results of our study are shown in Fig. 2. Their preliminary correlation is based on the regional seismic profiles. The Novocaspian, Mangyshlak, Khvalynian, and Khazarian complexes are distinguished in the studied sections.

The Novocaspian complex of the region overlies the upper Khvalynian and Mangyshlak sediments. It fills in and levels depressions in topography of the Mangyshlak time, being relief-forming on the other hand, because sediments of the complex form accumulative bodies on the bottom level of the Novocaspian time and recent rampart-like forms of different size. Depending on various factors, the complex thickness ranges from 0.2–0.3 m to 6.0–6.5 m. In the study region, the Novocaspian deposits are mostly represented by fine-grained sand with shell detritus beds and coquinas composed mainly of medium, coarse, and gravel-sized shells. The alternation of sandy and shelly stripes is observable directly on the bottom.

The structure and composition of the Novocaspian section is strongly different in the northern area 1 (Fig. 1), where the sea depth ranges from 7 to 9 m. The Novocaspian complex is 3.8–6.5 m thick here, and sandy-shelly grounds compose only its uppermost part. The most coarse-grained shelly deposits make up several rampart-like forms of latitudinal extension, which are up to 1.0–1.2 m high and situated at the altitudes of –34 to –36 m. Shells sampled from the base of one rampart are 860 years old. The altitude range, morphology, and available dates indicate that forms under consideration represent beach ridges of the Derbent regression time. The major part of the Novocaspian section in this area is mainly composed of mud intercalated with lens-like beds of fine-grained sand. Three sandy beds of different color, composition, and physical properties are recognized.

The Mangyshlak complex is represented in the study region by facies of greatly diverse genetic types, which reflect the consecutive southward movement of the deltaic zone and sea level lowering. Two genetic types distinguished in areas 1 and 2 are the river channel sediments and the intercalated lacustrine deposits.

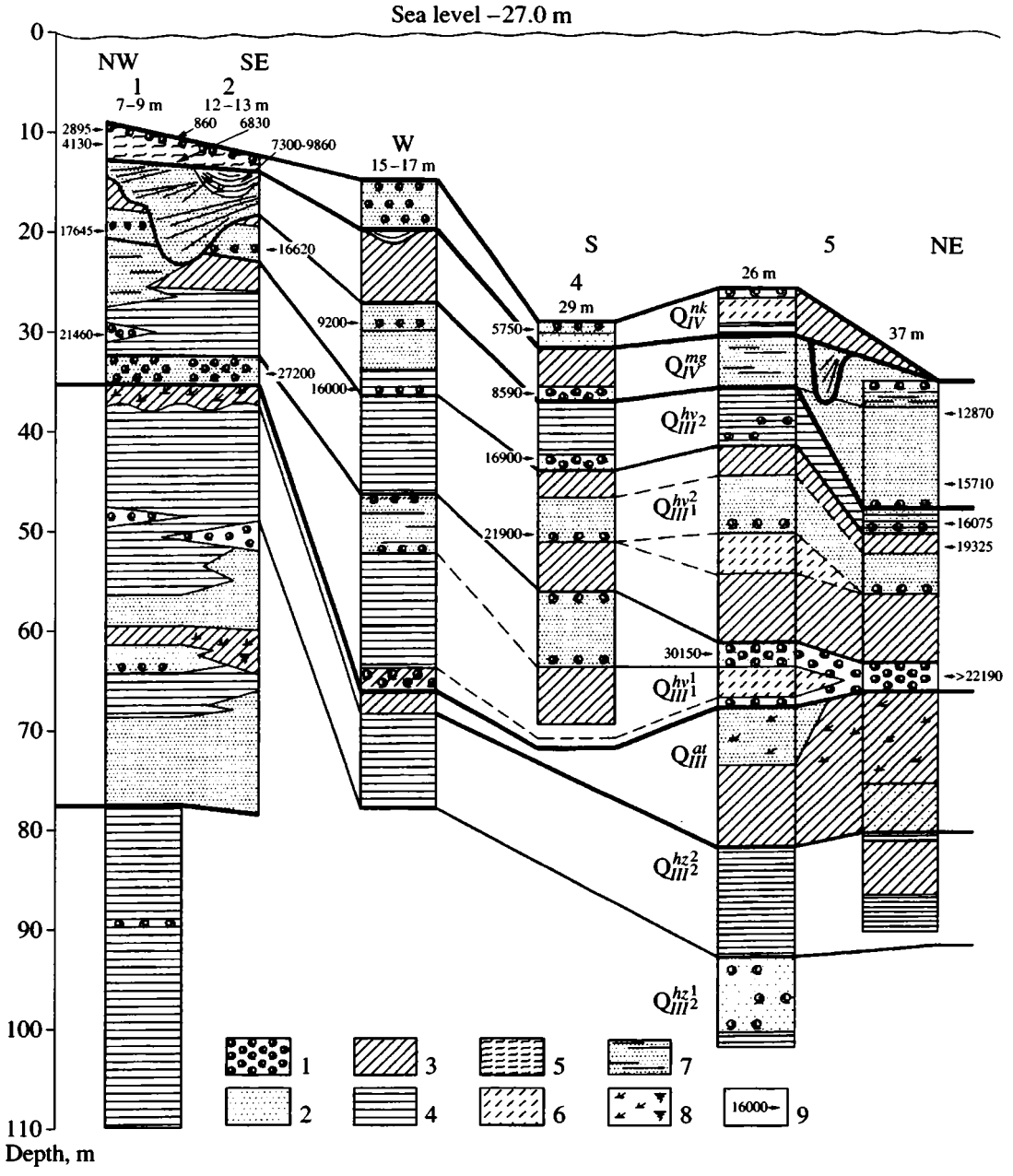

Fig. 2. Correlation scheme of studied sections: (1) coquina, coquina with sand; (2) fine-grained dusty sand; (3) dusty clayey deposits (clayey silt, loam); (4) clayey deposits; (5) clayey and loamy mud; (6) loamy sand; (7) sand with peat and organogenic mud intercalations; (8) plant detritus; (9) radiocarbon dates. Indices of stratigraphic horizons: (*nk*) Novocaspian; (*mg*) Mangyshlak; (*hv*) lower and upper (2) Khvalynian; (*hz*) Khazarian (upper). Shown at the top of columns are area numbers and sea depth values.

The former are represented by cross-bedded clayey fine-grained sand commonly bearing the fine plant detritus, scarcer shell fragments, "clay pellets," and by clayey silt (loam). Prevailing among lacustrine deposits are flowing clays with sapropel and peat beds, and with minor intercalations of dusty sand. These sediments fill in the buried depressions of sublatitudinal extension, which are 6 to 8 m deep. Four radiocarbon dates for sapropel and peat organic matter from the lacustrine complex range from 9300 to 9860 years. One date for peat corresponds to 7630 years.

In areas 3 and 4 located southward, the Mangyshlak complex is composed of dust clays and intercalated sand beds, which bear fragmented shells of fresh- and brackish-water mollusk. The total thickness of deposits is 7 to 9 m here. According to the seismic records, the section is of a bedded structure with the layers inclined and protruding south- and eastwards. The age of shells from the sand beds is analogous to that of the organic matter from lacustrine sediments and corresponds to 9230–8590 years. The deposits are dissected by the buried river channels of a variable depth.

Table 3. Facies environment and maximum depths in the studied regions of the Caspian Sea in the Quaternary

Event	Northern region (areas 1 and 2). Sea depth 8–13 m	Intermediate region (Area 3). Sea depth 15–17 m	Southern region (areas 4 and 5). Sea depth 25–37 m
Novocaspian	~860 yrs. ago, coast at –34–36 m; 2895 yrs. ago, shoals nearby the delta; 4130 yrs. ago, avandelta, shoals	Shoals; 6610 yrs. ago, shoals	Shoals; shoals, >5790 yrs. ago, avandelta
Mangyshlak regression	Land; 9800–9420 yrs. ago, land delta with lakes	Land; 9230 yrs. ago, avandelta	Land; 8540 yrs. ago, avandelta
Late Khvalynian	17645–16645 yrs. ago, shoals	Later the depth of 60–75 m; 16000 yrs. ago, shoals	Later the depth of 70–80 m; 16900 yrs. ago, shoals
Enotaevsk regression	Avandelta at –45–50 m	Shoals	Shoals
Early Khvalynian	21460 yrs. ago, shoals; Deep water ~100 m; 27200 yrs. ago, shoals	Deep water ~110 m	21090 yrs. ago, transition to shoals; Deep water ~130 m; 30150 yrs. ago, shoals
Atel' regression	Land	–	Lagoon deltaic shoals at about –100 m
Late Khazarian	Sea depth 70–80 m	Sea depth of about 100 m	Sea depth of about 130 m
Chernyi Yar–Astrakhan' regression	Shoals nearby the delta at about –90 m	–	Shoals at about –120 m
Early Khazarian	Deep water	Deep water	Deep water

According to results of seismic profiling, the largest incised channel is situated on the east of the study region near areas 2 and 5. Judging from the bottom topography, the channel extends from the Volga River valley down to the depth of 44–45 m (altitude of –73 to –74 m), being fringed on the south by a vast fan distinctly outlined by the isobath of 55 m. The incised channel is 8 to 12 km wide, 25 to 30 m deep, and almost completely filled with sediments. The filling sequence structure well recorded in seismic profiles indicates that deposits mostly accumulated during the ingressive stage at the time of sea-level rise. The age dates mentioned above for different facies types suggest that the Mangyshlak regression maximum happened between 6830 to 8540 years ago.

The Khvalynian complex is characterized by thickness increasing southward from 22–23 m in the northern area 1 to 30–35 m in areas 3 and 4. It is represented by beds of clay, silty clay, and silty fine-grained sand commonly bearing shelly detritus. The coquina bed at the complex base has sandy to clayey matrix and is up to 3–4 m thick in the northern areas. According to our data, in the Khvalynian complex section there is a distinct isochron of 16–16.62 ka that is confined on the north (areas 1 and 2) to the base of fine-grained sand with shelly detritus interlayers, and on the south (areas 3 and 4) it traces thin layers of shelly detritus included in clayey sediments. A sharp change in sediment colors is seen across the isochron. Silty to clayey deposits above the isochron are brown and brownish gray, while below sediments are gray to dark gray owing to presence of dispersed hydrotroilite. The mollusk assemblage from the upper part of sediments includes species characteristic of the upper Khvalynian deposits.

Below the isochron, the thickness of the Khvalynian complex increases southwards from 13–15 m in northern areas 1 and 2 to 30 m in areas 3 and 4. This part of the complex is composed of clayey and sandy beds bearing shelly detritus and resting on the basal coquina bed with sandy and clayey matrix, the maximum thickness of which is 3–4 m on the north. The species composition of mollusks may be considered as characteristic in general of the early Khvalynian time, although the basal coquina bed in area 2 contains forms more characteristic of the late Khazarian time. On the other hand, the obtained ^{14}C dates of 21.09, 27.20, and 30.15 ka correspond to the early Khvalynian time span. The two latter values characterize the basal coquina bed.

The Khazarian complex was penetrated by boreholes in the northern area 1 to the greatest depth. Its section is analogous here to those studied previously along the Volga-Caspian Channel that is situated to the northwest [5]. The upper Khazarian Subhorizon composed of sandy clayey deposits is recognized in the depth interval of 35 to 78 m below the sea bottom. It is underlain by the lower Khazarian Subhorizon composed of homogeneous semisolid and plastic clay. In areas 1 and 2, the upper half of the upper Khazarian sequence is represented by silty-clayey sediments and the lower one by beds of sand and clay. The basal sand bed is about 9 m thick. The peculiar characters indicative of sea shoaling or complete exsiccation are recorded near the top of the upper Khazarian Subhorizon. Indications of this kind are the abundance of plant detritus that accumulates usually along the delta front,

the occurrence of small plant roots, and holes left by burrowing organisms.

Late Quaternary Paleogeography

The discussed materials are useful for consideration of some paleogeographic aspects of the Caspian Sea history. All stages of the Caspian Sea level oscillations that were previously recognized in coastal sections and in the lower Volga region are recorded in the studied sections. The data considered above are presented in Table 3 to characterize in a generalized form the succession of major paleogeographic events in the study area. We indicated therewith the altitudes of shallow-water, deltaic, deltaic-coastal and lacustrine deposits, which record the sea level lowering during the regressive stages, and the inferred maximum depth of the sea in the northern Caspian region during the greatest transgressions. The depths are estimated based on published altitudes of the shore levels along the modern shoreline.

As it follows from Table 3, there were two periods of extreme lowering of the Caspian Sea level. The earliest one is recorded across the late Khazarian–early Khvalynian boundary. In the northern area, we found evidence of sediment accumulation in subaerial conditions at that time, and in the southern areas the lower Khvalynian sequence at the altitude of –100 m is underlain by the deposits bearing plant detritus and fragments of plant roots in their original position. We suggest that during that period corresponding to the so-called Atel' regression, the sea level dropped to the altitude of –100 m (about –70 m below the current level) and was even greater probably. The second large, so-called Mangyshlak regression took place in the terminal late Pleistocene– early Holocene time. The sea level was at that time at the altitude of –73 to –75 m, or about –45 m below the present-day level.

The shallow-water deposits of the coastal and deltaic-coastal types are recorded sections between the lower and upper Khazarian marine sediments at the altitude of about –90 m, and between the lower and upper Khvalynian marine sediments at the altitude of –45–50 m. These two events correspond to the Chernyi Yar (or Astrakhan) and Enotaevsk regressions, respectively.

During transgressions, the depth in northern areas of the Caspian sea was as great as 100–130 m in the early and as 70 m in the late Khvalynian time.

CONCLUSIONS

The comprehensive study of continuous sections of the Upper Quaternary sediments recovered by boreholes from depth range of 7 to 37 m and the results of continuous seismic profiling elucidated new facts of the Caspian Sea geologic history.

(1) Biostratigraphic analysis enabled discrimination of successive lower and upper Khazarian, lower and upper Khvalynian, Mangyshlak, and Novocaspian horizons. The mollusk assemblages from the deep-water deposits of the early and late Khvalynian time are first distinguished. They include characteristic species *Didacna parallella* (Bog.), *D. subcatillus* Andrus., and scarcer *D. praetrigonoides* Nal. and *D. cristata* (Bog.) of the early Khvalynian time and *D. praetrigonoides* Nal., *D. cristata* (Bog.), and rare *D. parallella* (Bog.) and *D. subcatillus* Andrus. of the late Khvalynian time. The assemblages slightly changing in composition over the study region substantiate a reliable correlation of the sections. They differ from concurrent assemblages of the northern Volga region, where the corresponding shallow-water deposits contain numerous *D. protracta* (Eichw.), *D. trigonoides* Pall. and scarcer *D. parallella* (Bog.) of the early Khvalynian time and *D. praetrigonoides* Nal. (Svitoch and Yanina, 1997).

(2) About 40 radiocarbon dates are first obtained for sediments recovered in the Caspian basin. These results clarify controversial views on the age and accumulation time spans of the distinguished horizons, which were discussed in a previous publication [2]. We established that the Novocaspian, Mangyshlak, upper and lower Khvalynian sediments were deposited in the time spans of 0–7, 7–9, 9–16, and 17 to >30 ka, respectively.

(3) The facies composition and attitudes of the Late Quaternary deposits are used to evaluate oscillations of the Caspian Sea level. We substantiate for the first time that the sea level lowered up to –60, –70, –20, and –45 to –50 m below the present-day stand during the mid-Khazarian, Atel', Enotaevsk, and Mangyshlak regressions.

Reviewers Yu.A. Lavrushin and M.N. Alekseev

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